THE GEOCHEMICAL EVOLUTION OF FIJI

by

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Analyses and arguments presented in this thesis are my own unless otherwise indicated.

James B. Gill

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All analytical data reported herein are results of my own work except those in Jakes and Gill (1970) which are Jakes', the Solomon Islands analyses in Gill and Compston (Appendix 3) which are Compston's, and the New Hebrides analyses in Gill and Gordon (Appendix 3) which are Gorton's. Ar extractions and mass analyses were joint projects with Dr. I. McDougall and Mr. Z. Roksandik. I wrote each paper in Appendix 3 involving joint authorship.

Computer computations were used in all data reductions and in calculation of fractionation and partial melting effects. Programs not acknowledged are my own.

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ABSTRACT

The geochemistry of igneous rocks from Fiji is relevant to such complex issues as the tectonic history of Fiji and eastern Melanesia, the genesis of magma in island arcs, and the origin of continents. To these ends, new chemical analyses of 172 Fijian rocks are presented, each of which include data for major and fifteen trace elements. The isotopic composition of Sr in 58 of these samples was measured; concentrations of thirteen rare earth elements, Hf, and Sn were also determined for 27 of them. Five K-Ar age determinations were made and about two dozen partial analyses of samples from Java, New Britain, the New Hebrides, and Tonga are also included.

Fiji appears to be an incipient rather than vestigial continent which grew largely in response to the northward movement of the Australian relative to the Pacific plate during the Tertiary. The earliest record of its history is recorded in the basal rocks of Viti Levu. Study of these rocks suggests that the Lau-Colville Ridge (of which Fijian islands are pinnacles) evolved between the early Tertiary and the early Middle Miocene as a largely submarine island arc through igneous activity of the island arc tholeiitic series type. One interpretation consistent with my data and with most of what is known about the regional geochronology is as follows. In Late Miocene time a subduction site lay east and north of the Lau-Colville Ridge, judging from spatial variations in the geochemistry of volcanic rocks of this apparent age. Gradually the subduction site moved to its present position east of Tonga and left the Lau-Colville Ridge behind, with an enlarging Lau Basin in between. Shoshonitic volcanism accompanied this transition in tectonic style. Subsequent volcanism became predominantly basic and, eventually, more typical of that associated with intra-oceanic islands.

There are relatively consistent spatial variations in the geochemistry of magmas in modern island arcs which are summarized in Figures 3.1 to 3.3. If any of these magmas rise directly from their subjacent Benioff zones and are primary partial melts of underthrust lithosphere, they must represent liquids in equilibrium with eclogite. Garnet and clinopyroxene were found by T. Green (1972) to be high-pressure near-liquidus phases of a Late Miocene Namosi andesite. Major and trace element and isotopic data were therefore combined with Green's experimental results to evaluate in detail the possible role of

underthrust ocean floor basalt and sediment in the genesis of the Namosi Andesite suite. Although the problem is insufficiently constrained, rocks of this suite seem unlikely to be primary partial melts or high-pressure fractionation products from partial melts of oceanic crust, if that crust has the composition of "average" fresh or altered ocean floor basalt with or without contamination by oceanic clay or carbonate. Similar reasoning suggests that no other Fijian suite nor any known island arc suite is geochemically compatible with derivation involving partial melting or crystal fractionation of eclogite. Partial melting at the Benioff zone of amphibolite with the composition of ocean floor basalt requires very considerable shear-strain heat sources and does not explain the geographic position of island arc volcanoes or the geochemistry of their magmas.

The composition of Fijian volcanic rocks is equally inconsistent with them having been primary partial melts of the upper mantle or high-pressure fractionation products thereof. A proposal by which to explain their composition and spatial variations in island arc magma geochemistry is offered. It involves mixing of the low-melting fraction of both underthrust lithosphere and overlying upper mantle, differences in the depth and degree of fusion of the latter, and low-pressure differentiation of resulting magmas under conditions in which partial pressures of water and oxygen increase relatively consistently with distance from the trench. The compositions of shoshonitic and other Pliocene to Recent mafic volcanics in Fiji are thought to reflect migration of the subduction zone and therefore gradual removal of the source of primary intermediate to acidic magmas, water, and heat. The amount and stability of accessory minerals such as kaersutite and those of the perovskite group in the upper mantle may change as this migration proceeds.

An estimated average composition of crust produced by pre-Pliocene volcanism in Fiji is presented as an example of the contribution of island arc development to crustal evolution. Crust with the composition of that found in continental shield areas cannot have originated solely through an earlier but geochemically similar process. Some other source of at least Rb, U, Th, and Pb is necessary.

Ways in which this thesis might be applied in Fiji to environmental health, ore exploration, and archaeology are sketched in a concluding note.

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Chapter 1. Introduction

Fiji has long been thought an incipient or vestigial continent. This particular study of it was presaged by the agreement in 1965 of R. Glover, then Director of the Fiji Geological Survey Department, to contribute two Fijian samples to S.R. Taylor who was studying the geochemistry of andesites. Analyses of one of these and a subsequently donated sample, reported by Taylor et al. in 1969, were sufficiently close to Taylor's "average andesite" and "average crust" that Fiji appeared to be a promising site for investigating the geochemistry of andesites and continental evolution. Glover's 1965 invitation for an ANU student to work on a Fijian topic resulted in this study, conducted between November, 1968 and May, 1972. What was known in 1968 about the geochemistry and geological history of Fiji was summarized by Rodda (1967 and 1969) and Dickinson (1967), and defines the context of this thesis. As my work progressed, the existence and significance of spatial and temporal geochemical variations were recognized which, in turn, led me to attempt a unified interpretation of the geochemical and tectonic evolution of Fiji and its environment. The articles written prior to the thesis itself and included in Appendix 3 are integral to it and should be read at the outset because many points raised in them are not developed elsewhere in or are used as background for the thesis. In any discrepancies of data or interpretation between earlier articles and thesis text, the latter takes precedence.

The two central issues of this dissertation --the tectonic history of Melanesia and the genesis of island arc magmas-- have been geological conundrums for a century. My purpose is to use the petrochemistry of Fiji's igneous rocks to unite both within the family of ideas comprising plate tectonic theory. To do so requires faith that the general formulations of that theory by, for example, Vine and Hess (1970), LePichon (1968), Morgan (1968), and Isacks, Oliver, and Sykes (1968), approximate what is going on now and has been going on throughout at least the Tertiary. While many things qualify ones confidence in the simplicity of these formulations, they remain the most elegant, most satisfactorily comprehensive, and most heuristic paradigms in terms of which to operate. I shall therefore take as fact that lithosphere is consumed at or near the site of ocean trenches despite the criticisms of Carey (1970), Rickard (1969) and Hatherton (1971).

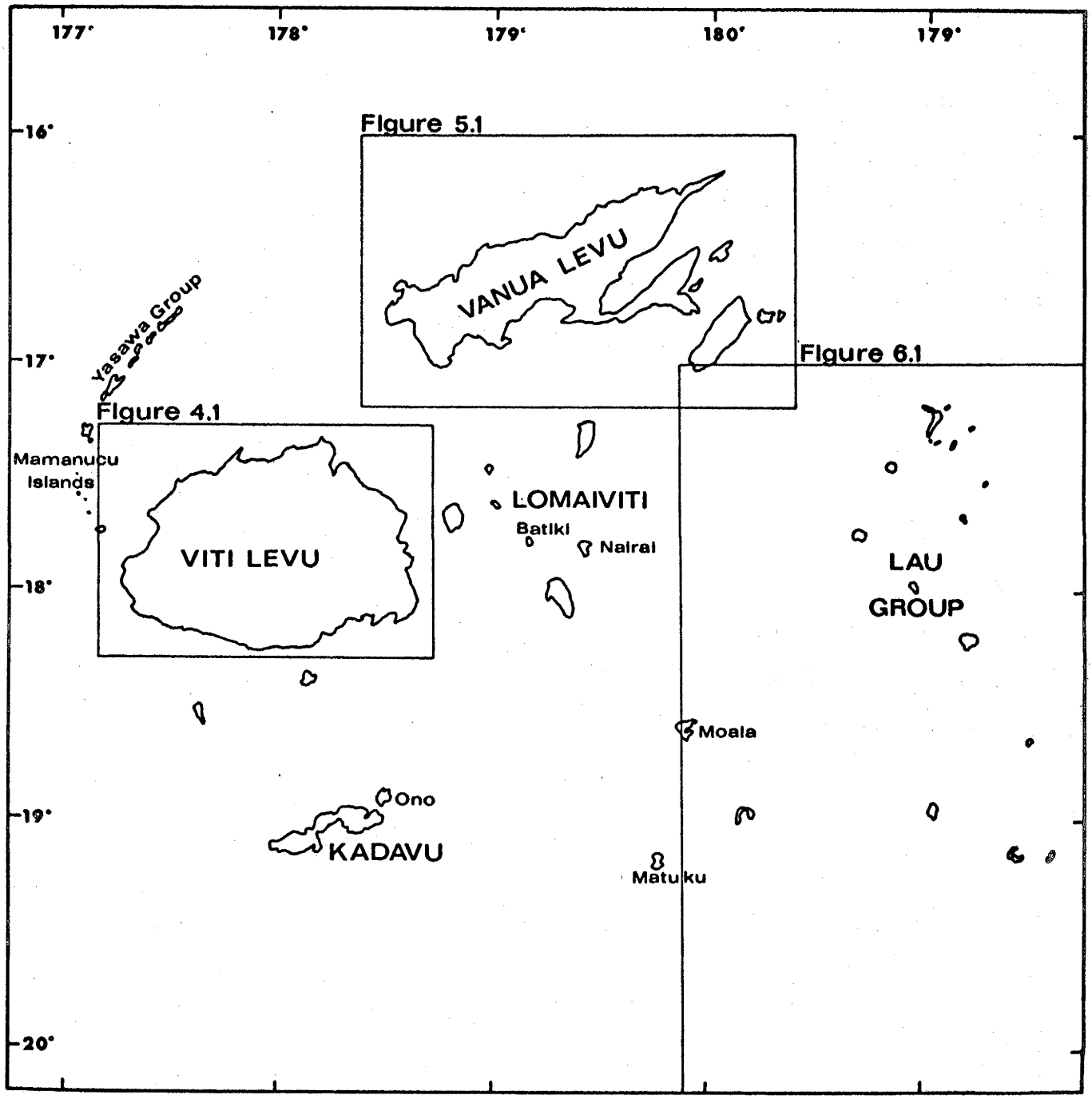

Figure 1.1 Map of Fiji. Fiji's position in the southwest Pacific is illustrated in Figure 1 of Gill (1970). Bathymetric relationships are shown in Figure 6.1 for the Lau Group and for the entire region in several figures in Gill and Gorton (Appendix 3). Only the Lomaiviti islands from which I report analyzed specimens are identified.

Even within such theory my two central issues remain conundrums. Only the obvious is indisputable: that underthrusting of lithosphere is in some way a cause of island arc volcanism, and that Melanesian tectonism is in some way the result of interactions between the westward-moving Pacific and northward-moving Australian plates. Beyond that little is firm. What is needed in such complexity is insight; what is common is either provincial pessimism rendered silent by the complexity, or imaginative speculation which treats complexity as licence. By temperament I probably err in the latter direction, and some of my conclusions may be criticized as premature or, at best, non-unique. Nevertheless, at least one goal of education is an ability to decide what is likely when little is certain and it is that toward which I have striven.

Discoveries of the last decade have emphasized the intimate and causal relationship(s) between igneous activity and tectonism (e.g. D. Green, 1972). My thesis is rooted in this premise. It is a look at the evolution of Fiji through the window of its igneous rocks, convinced that magmatism is not a fortuitous accident but is, instead, related to large scale tectonic events.

1.1 Scope

This thesis is, as are most, multi-faceted. I view it as: (1) a contribution to the descriptive geochemistry of island arcs that is relatively comprehensive both in the amount of its analytical data per sample and in the spatial and temporal distribution of those samples. In particular it follows the history of one of the few arcs in which one can trace development from inception as a submarine ridge to old age as a "third arc". This has involved analyzing the concentration of 25 to 50 elements in almost 200 rocks having an age range of at least 40 m.y. and covering some 17,000 km^2 of land surface spread over more than a dozen islands (Figure 1.1). It is (2) an attempt to explain the ambiguous tectonic evolution of eastern Melanesia during the Tertiary using the volcanic rocks as principal guidelines for reconstruction in terms of plate tectonics; and (3) an investigation of island arc magma genesis by reference to the geochemistry of Fijian volcanic rocks.

It is equally important to say what a thesis is not. Mine is not a detailed study of: (1) one volcano, one island, or one formation involving field mapping, extensive petrographic study, optical or

chemical mineralogical data, and tight stratigraphic control; (2) factors affecting the distribution of some trace elements; or (3) specifically Fijian problems of stratigraphic correlation, inference of paleo-environments, or real genetic relationships between actual rock samples. Most details having only local significance are omitted, but not all because I have no other way to record them. Some, especially those involving stratigraphic identification of samples, are necessary inclusions.

The breadth of my objectives was accepted for both educational and scientific reasons. Only by such reconnaissance study was it possible to document the spatial and temporal variations in geochemistry which witness Fiji's tectonic history and which constrain theories of magma genesis there. Such breadth has forced many choices while conducting my work and writing this thesis. It has not been possible to do or say all that could be done and said. In particular, I have not thoroughly explored all possible genetic implications of my data for the rock suites discussed in chapters 10 to 12. Breadth also requires empirical generalizations which, in turn, entail simplifications. I hope mine do not distort field or analytical complexity so much that my overall arguments are invalidated.

Because Fiji is not part of an active island arc today, mine cannot be an actualistic study of island arc magma genesis. Indeed, the possible atypicalness of Fiji is a recurring phantom which has haunted the thesis. Thus the arguments presented by Gill and Gorton (Appendix 3) are not only significant as a possible interpretation of the geological history of Fiji; they also constitute the basis upon which my data gain relevance to the larger issues of island arc magma genesis. The tectonic history and geochemical evolution of Fiji are opposite sides of the same coin.

1.2 Conventions

All analyses reported herein have been calculated on a loss-free (anhydrous) basis for major and trace elements. Loss on ignition and original weight per cent totals are given in parentheses. FeO contents were not always determined due to scepticism about their value.

Because iron oxidizes during and after eruption, many either correct Fe_2O_3/FeO ratios to the lowest value they measure for a given suite, fix a maximum Fe_2O_3 content and recalculate anything above it

to FeO (Coombs, 1963; Kay et al., 1970), calculate Fe_2O_3 by taking per cent TiO_2 + 1.5 (Irvine and Baragar, 1971), or fix a constant or maximum Fe_2O_3/FeO ratio (Chayes, 1966). For simplicity I have set $Fe_2O_3/FeO = 0.25$ in calculations of CIPW norms and $Mg/Mg+Fe^{+2}$ ratios for this thesis, unless otherwise noted, even though measured ratios exceed this value and vary within suites. This procedure may overestimate FeO contents and thereby decrease the magnetite content and degree of silica saturation in normative assemblages, make normative mafic minerals more ferrous, and lower $Mg/Mg+Fe^{+2}$ ratios. Fe_2O_3 contents estimated in this way generally agree with or exceed those calculated following the procedure of Irvine and Baragar (1971).

All 87Sr/86Sr ratios reported as new or quoted in this thesis have been normalized to 0.7080 for E&A $SrCO_3$.

Data presentation generally follows the format proposed by Taylor (1965). No more than three figures are significant in the elemental abundances and ratios, except for 87Sr/86Sr ratios where all four are. REE are displayed in figures as "rock" or "melt enrichments". These represent rock or melt concentrations relative to the enrichment factors listed in Table 4.2, which are approximately the REE concentrations in ordinary chondrites. $Mg/Mg+Fe^{+2}$ is used as an atomic ratio.

1.3 Nomenclature

Nomenclature is the Pandora's box of petrology. In this thesis I have adopted a simplified system based entirely on chemical composition rather than modal minerology. This system has evolved gradually through frequent arguments between members of the Geology and Geophysics Departments, ANU. Four principal rock series are recognized in island arc environments: island arc tholeiitic, calc-alkaline, high-K calc-alkaline, and shoshonitic. None of these are mutually exclusive categories and the nomenclature convention is merely an attempt to subdivide usefully the complex variety of island arc rocks. Generalized geochemical descriptions are given for these rock series in Table 1.1. While K_2O-SiO_2 variations are thought to be the most useful distinguishing criteria, all aspects of composition should be considered when selecting semantic pigeon-holes.

The names chosen for these series follow the suggestions of Taylor et al. (1969), Joplin (1968), and Jakes and Gill (1970) by taking more note of ordinary language and use than of etymological

Figure 1.2 Non-basalt nomenclature.
( ) indicates approximate maximum $\%K_2O$ at 60% SiO_2

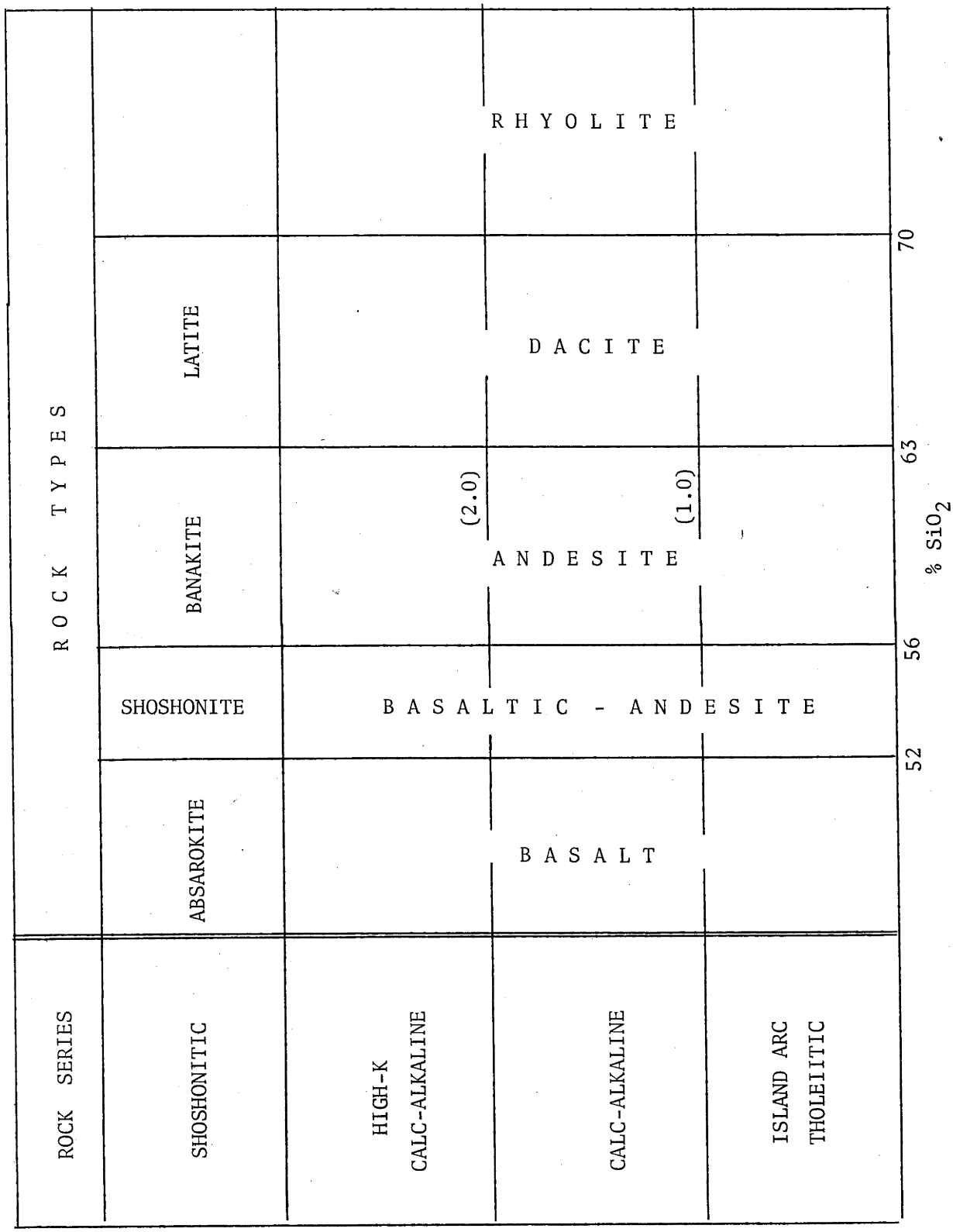

Figure 1.3 Basalt nomenclature*

$\frac{An}{An+Ab}$	DI**	hy>2%	hyp<2%; ne>0.5%	0.5%<ne <5%	ne>5%
>44	<65	Tholeiite Basalt (olivine- or quartz-)	Olivine Basalt	Alkali Olivine Basalt	Basanite
30-44	<65			Hawaiite	Ne-Hawaiite
10-30	<65			Mugearite	Ne-Mugearite
	65-75			Benmoreite	Ne-Benmoreite
	>75			Trachyte	Phonolite

* Used when $Na_2O/K_2O > 2$.

** DI = Qz + Ab + Or + Ne + Lc.

Table 1.1. Geochemical summary of island arc rock series. Minor and trace element data in columns A,B, and C apply to andesitic members.

	A. Arc Tholeiitic	B. Calc-alkaline	C. High-K Calc-alkaline	D. Shoshonitic
SiO_2 range	45-80%	53-70%	53-70%	48-68%
SiO_2 mode	basaltic-andesite	andesite	andesite	basalt
TiO_2	1%	0.7-0.8%	0.7%	0.7%
Al_2O_3	14-19%	16-19%	16-18%	13-20%
Max. $FeO+Fe_2O_3$) at 5% MgO	10-12%	< 9%	< 8%	9%
K_2O	< 1%	1-2%	> 2%	2-7%
P_2O_5	0.2%	0.3%	>0.3%	0.4%
Rb	≤10ppm	20-30	>50	60-100
Sr	100-200	300-600	>600	>1000
Ba	50-150	200-400	>400	600-900
Pb	≤2	5	7	10
La	≤6	9-12	>12	5-15
Yb	>2	1-2	1-2	1-2
Y	30-40	20-25	20	15-25
Th	0.5	2	4-6	1-2
U	0.3	0.7	2	0.5-1.0
Na_2O/K_2O	>5	2-3	1.5	1.0
K/Rb	>700	500	300-400	300-500
Rb/Sr	<0.03	0.03-.06	>0.06	0.04-.06
Ba/Rb	25	20	15	10-15
Th/U	1-2	3-4	2-3	2
La/Yb	<3	6-8	>10	4-8

consistency and thereby overlook the disapprobations of Chayes (1966) and Nicholls and Carmichael (1969).

Within each rock series I find it useful to distinguish further on the basis of SiO_2 content, as indicated in Figure 1.2. I shall also use the adjectives "basic" and "mafic" to refer to rocks with <52% SiO_2; "intermediate" to denote rocks with 52 to 63% SiO_2; and "felsic" or "acidic" to denote rocks with >63% SiO_2.

When describing predominantly basaltic suites I have adopted the nomenclature used by Irving (1972) after Coombs and Wilkinson (1969) and Green and Ringwood (1967), which is set out in Figure 1.3.

1.4 Orthography, abbreviations, and stylistic format

I have used the "Fijian" spelling system for Fijian place names, largely for brevity and to conform to local Fijian custom. In doing so I follow Derrick's (1965) example. When saying words spelled in this manner, b is pronounced mb, c is pronounced th, d is pronounced nd, g is pronounced ng (as in sing), and q is pronounced ngg (as in hunger). These phonetic equivalents have been adopted in most publications designed for overseas reading, including those of the Fiji Geological Survey Department, and I have retained their spelling of rock unit names. Thus, rocks of the Mba Basaltic Group will crop out near the town I shall spell Ba, and the Korombasanga Volcanics crop out on the hill I shall spell Korobasaqa.

Commonly used abbreviations are REE for rare earth element(s), LOI for loss on ignition, and FGSD for Fiji Geological Survey Department. Other abbreviations are used in certain sections where they occur repeatedly and are defined before usage (e.g. EJVB for East Japan Volcanic Belt).

The thesis format follows the conventions set out in the "MLA Style Sheet" except for reference citations which adopt the procedures of Geochimica et Cosmochimica Acta.

1.5 Sample representativity

Representativity must always be questioned, especially in a study of this scope. In collecting samples, I sought to obtain material representative of the unweathered, least-altered igneous rocks of each mappable rock unit in Fiji containing such. They are not a collection of the extremes of rock types available and may, therefore, be vaguely "typical" of their respective rock units although I have

insufficient field experience in Fiji to confirm this.

Watkins et al. (1970), Hart et al. (1971), and others have documented considerable geochemical variability within single lava flows, particularly for those elements excluded from crystallizing minerals. I made no such study but did collect and analyze multiple samples from single outcrops. Table 1.2 presents these comparative results. The fluctuations represent a combination of outcrop variability and analytical error and therefore define a limit below which no genetic significance other than crystallization effects should be attached to differences between analyses of samples.

When pillow lavas were sampled, specimens were taken from cores of the largest accessible pillows.

1.6 Neogene time-scale

Regional biostratigraphy in Fiji as elsewhere in the Indo-Pacific region is based on tropical, larger foraminiferal fauna and the East India letter stages. In FGSD publications these stages are usually used interchangeably with Lyellian epoch names to which absolute time significance is attached.

In Figure 1.4 I present the correlation between Indonesian letter stages, Lyellian epochs, planktonic foraminiferal zones, and time (in 10^6 years) which I have adopted. This follows the correlations proposed by Adams (1970), Berggren (1969), and Page and McDougall (1970), and differs from Blow's (1969, Figure 20) only in calling the f_3 stage Late rather than Middle Miocene. Figure 1.4 also summarizes the stratigraphic correlations adopted in chapters 4 through 8.

1.7 Thesis organization

Chapters two and three continue the introduction by reviewing the geological characteristics and tectonic history of Melanesia (chapter two) and spatio-temporal variations of geochemistry in the world's island arcs (chapter three). Both are designed to provide information with which to assess Fiji's former relationship to plate behaviour in the southwest Pacific.

My analyses of rocks from Viti Levu (chapter four), Vanua Levu (chapter five), the Lau Islands (chapter six), and Taveuni, Kadavu, the Lomaiviti, and Mamanucu Islands (chapter seven) are then presented. For each island or island group I briefly summarize the

Table 1.2. Replicate analyses from single outcrops.

*Clast from agglomerate

	50.	51.	61.	62	367	60	913	63	863	71*	865*	896a	896b	48	386	66	X88	PJ6	PJ18
SiO_2	73.31	73.45	47.68	47.98	49.20	59.22	58.55	52.03	51.80	50.08	50.27	51.66	52.05	57.46	56.99	60.09	60.17	59.62	59.62
TiO_2	0.51	0.51	1.13	1.18	1.07	1.08	1.05	0.56	0.56	1.01	0.69	0.87	0.79	0.64	0.64	0.69	0.71	0.67	0.70
Al_2O_3	13.13	13.07	18.86	19.41	18.51	15.13	15.01	17.56	17.11	16.03	15.19	16.46	16.09	19.57	19.61	16.93	17.26	17.12	17.24
Fe_2O_3	1.59	1.89	7.84	9.39	11.12	11.29	11.82	7.55	8.75	11.06	10.30	10.74	9.99	7.68	7.65	3.70	2.98	3.83	3.94
FeO	2.09	1.81	3.86	2.82												2.64	2.93	2.60	2.59
MnO	0.12	0.12	0.24	0.22	0.18	0.15	0.17	0.18	0.17	0.20	0.20	0.20	0.17	0.15	0.15	0.13	0.26	0.13	0.15
MgO	0.78	0.68	6.23	5.35	5.52	3.50	3.35	3.46	4.38	5.85	5.18	4.94	5.49	2.30	2.50	3.12	3.18	3.44	3.28
CaO	2.65	2.70	10.06	9.59	10.37	6.17	6.24	9.25	9.22	10.21	11.71	9.54	9.87	8.74	8.54	7.20	7.32	7.03	7.16
Na_2O	4.78	4.54	2.82	2.95	2.77	3.32	3.26	5.11	3.70	2.93	2.86	2.80	2.76	3.02	2.97	4.02	3.79	3.97	3.84
K_2O	0.91	1.11	0.97	0.79	0.96	0.30	0.41	4.12	3.76	2.75	2.81	2.43	2.44	0.79	0.84	1.28	1.26	1.38	1.28
P_2O_5	0.12	0.11	0.30	0.32	0.31	0.13	0.14	0.45	0.45	0.39	0.79	0.36	0.35	0.10	0.10	0.20	0.14	0.20	0.20
(LOI)	(1.79)	(1.83)	(4.32)	(4.67)	(1.50)	(4.24)	(3.86)	(3.33)	(4.86)	(1.41)	(5.95)	(1.54)	(1.81)	(2.21)	(2.33)	(0.55)	(0.67)	(0.85)	(0.64)
Rb	6.9	9.1	8.3	5.0	12.	2.2	7.4	73.	70.	30.	55.	45.		7.2	7.2	16.	22.	23.	19.
Sr	118.	12.1	619.	522.	583.	134.	139.	1600.	1710.	1300.	1210.	747.		172.	174.	487.	449.	517.	506.
Ba	222.	285.	169.	190.	137.	74.	62.	698.	632.	462.	956.	398.	395.	112.	108.	467.	333.	562.	537.
Pb			3.	3.	2.	2.	2.	11.	6.	7.	9.	6.		4.	2.	4.	2.		
Y	44.	45.	23.	24.	22.	36.	38.	13.	21.	23.	29.	27.	26.	29.	27.	17.	20.	18.	18.
Th			.63	.80	.88	.20	.10	2.3	2.2	1.0	2.7	.53				2.0	1.9	1.8	1.3
U	.20	.20	.24	.17	.31	.10	.10	1.2	.89	.50	1.8	.21				.72	.76	.76	
Zr	72.	74.	65.	72.	61.	71.	69.	67.	45.	65.	107.	48.		56.	53.	112.	116.		
Ni			20.	12.	10.			6.	13.	17.	25.	26.	30.			11.	8.	9.	9.
Co		3.	40.	34.	35.	28.	32.	21.	25.	32.	39.	39.	34.	18.	17.	16.	16.		
Cr		2.	32.	7.	8.	2.	5.	3.	14.	9.	35.	53.	93.		3.	31.	36.	46.	42.
Sc	16.	18.	39.	36.	29.	37.	43.	15.	23.	29.	33.	37.	34.	29.	21.	14.	18.		
V	9.	9.	304.	265.	204.	200.	247.	340.	274.	355.	349.	256.	231.	154.	121.	172.	162.	196.	159.
Ca		2.	91.	83.	129.	81.	88.	185.	203.	101.	143.	164.	134.	32.	36.	42.	29.	44.	30.

EPOCHS		Indonesian Letter Stages	Time (in 10^6 yrs)	Planktonic Zones	Adopted Fiji Equivalents
PLIOCENE		h	5.5	N.18-21	Third Period Viti Levu; Mbua; Taveuni; Lomaiviti.
MIOCENE	Late	g	9.0	N.15-18	Namosi Andesites; Mathuandrove Super-Group; Lau Volcanics; Kadavu Group (?).
		f_3	12.5	N.13-15	
	Middle	f_{1-2}	15.	N.9-12	
	Early	e_{upper}	22.5	N.4-8	First Period Viti Levu
OLIGOCENE	Late	e_{lower}		P.18-N.3	
	Middle	d			
	Early	c	36.		
EOCENE	Late	b	45.	P.15-17	

Figure 1.4 Time equivalent designations and a Fijian stratigraphic column

available geological, petrological, and geochemical information, aware that I am simplifying frequently complex geology to which I have had limited personal exposure. The presentation is in terms of the largest stratigraphic units within which igneous rocks are likely to be genetically related and/or to have formed during a single configuration of relative plate movements.

These results are summarized in chapter eight and Appendix Three (manuscript by Gill and Gorton) where I present a unified hypothesis for the volcanic and tectonic evolution of the Fiji area in a manner consistent with plate tectonics and the information presented in chapters two and three.

In chapter nine I review issues relevant to magma genesis in island arc environments and then use my analyses of Fijian rocks to constrain ideas about their source regions and genetic histories. I do so by considering the genesis of Fijian calc-alkaline and high-K calc-alkaline series (chapter ten), island arc tholeiitic series (chapter eleven), and shoshonite-basalt suites (chapter twelve).

A few comments about the composition of Fiji as a Tertiary continent and a brief summary are given in chapter thirteen. Analytical techniques and reliability are discussed in the first appendix; geographic and petrographic information for all analyzed samples are provided in the second; reprints and preprints of published articles which are submitted as part of the thesis constitute the third; and CIPW norms for all analyzed basalts (as defined in this chapter) are given in the final appendix.

Chapter 2. The Tectonic Context: geological complexity of Melanesia

2.1 Introduction

Fiji has often been considered a tectonic enigma (e.g. Coleman, 1966, p.1251; Hayes and Ewing, 1970, p.44). To the east of Fiji the bathymetry, distribution of earthquake epicenters and volcanoes, and earthquake focal mechanism solutions all show that the Pacific is being thrust westward under the Australian plate at the present time (Oliver and Isacks, 1967; Isacks et al., 1969; Isacks and Molnar, 1971; Karig, 1970; Mitronovas and Isacks, 1971). To the west of Fiji the same criteria indicate northward underthrusting of the Australian beneath the Pacific plate (Santo, 1970; Jakes and White, 1969; Isacks and Molnar, 1971; Denham, 1969; Johnson et al., 1971; Johnson and Molnar, 1972).

Has Fiji always been in the transition zone between these regimes? If so, its history may be unique and without similarity to island arcs which evolved through more normal subduction processes. Malahoff (preprint) explicitly, and Chase (1972) and others implicitly adopt this view. If correct it means Fiji may be a poor place to study island arc volcanism because of its atypicalness. Or did Fiji avoid tectonic schizophrenia by belonging to one of these two subduction regimes and, if so, to which and when did its present isolation commence?

There are at least three ways to approach this problem. First, Fiji itself can be examined and its history compared to the evolutionary pattern of "normal" island arcs which has been summarized by, for example, Dickinson (1970; 1971b,c), Mitchell and Reading (1969, 1971), or Dewey and Bird (1970). Although Fiji's sedimentary and igneous lithologies, facies relationships, and evolutionary pattern (Rodda, 1967; Dickinson, 1967, 1968a) suggest an island arc environment, they are also consistent with suigeneric development in its present non-arc position. In chapter three I will summarize what is known about variations in the geochemistry of volcanic rocks from modern island arcs and in chapters four to eight (see also Gill, 1970, and Gill and Gorton, Appendix 3) will describe the geochemistry of Fiji and compare it to this norm.

A second approach is taken in the following section where Fiji's modern environment is examined so as to assess its present and former

relationship to the subduction regimes between which it now lies. A final constraint is added in section 2.3 by considering the long-term history of the eastern Australian plate to which Fiji belongs.

2.2 Bathymetric and structural summary of the modern Fiji area

Political Fiji, excluding Rotuma, is merely the above-sea-level portion of a 2500 km long, golf-club shaped ridge (the Lau-Colville Ridge) extending from 15°S to the North Island of New Zealand (Figure 1.1). There is neither subaerial nor submarine evidence for earlier discontinuity of ridge elements as required by Chase (1972, Figures 11c,d). Volcanic rocks within the exposed basement of Vanua Levu and the Lau Group are of similar age and composition (chapters five and six), so there is no a priori reason to suspect that their present proximity is fortuitous.

Fortunately, the sea floor surrounding Fiji and the region's seismicity have been fairly well studied by several groups. I was able to augment my reading by briefly inspecting the Lau Basin between 15 and 21°S first-hand during the 1971 Antipodes cruise of the R/V "Melville". Salient characteristics summarized from these various sources are as follows.

(1) The crust of the Lau-Colville, New Hebrides and Tonga-Kermadec Ridges appears to be 20 to 30 km thick (Shor et al., 1971; Officer, 1955; Hunkins and Kuo, 1965; Solomon and Biehler, 1969; Malahoff and Woolard, 1969; Malahoff, in press). The crust beneath Viti Levu and Vanua Levu, Fiji, is at the thickest end of this range (Robertson, 1967; Worthington, pers. comm., 1971). Published seismic refraction work is not adequate to provide reliable velocity profiles (compare, for example, the sections of Horibe (ed., 1970, Figure 72) and Shor et al. (1971, Figure 3), but ridge basements seem to have velocities between 7.0 to 7.5 km/sec.

(2) The surrounding ocean basins have thinner crust, though slightly thicker than standard oceanic profiles. The South Fiji Basin has thicker crust, more sediment, lower heat flow, and is deeper than the Lau Basin or Fiji Plateau (Shor et al., 1971; Karig, 1970, 1971b; Solomon and Biehler, 1969; Sutton et al., 1970). The latter may also be underlain by a layer with $7\frac{1}{2}$ km/sec velocities (Sutton et al., 1970, Figure 10).

(3) The Melanesian Border Plateau, which forms the area's northern boundary at 13-14°S, seems to be a drowned intra-oceanic

volcanic archipelago of Cretaceous age which was capped by young volcanism while passing over a mantle "hot-spot" presently located beneath Samoa (Fairbridge, 1961; Fairbridge and Stewart, 1960; Chase, 1972).

(4) The Lau Basin and Fiji Plateau are young (<10 m.y. old) marginal seas. Their youthfulness is suggested by their thin (<100 m) or non-existent sediment veneer, rugged topography, and very high (up to 20 $\mu cal/cm^2/sec$) heat flow (Sclater and Menard, 1967; Karig, 1970; Sclater et al., 1972; Chase, 1972; Malahoff, 1970; Kurentsova and Shreyder, 1971). Fossils older than Pliocene are rare (Chase, 1972; preliminary results of the 1971 Antipodes cruise in the Lau Basin). Youthfulness is also implicit in their shallowness (2 to 3 km), following the arguments of Sclater et al. (1971) and Karig (1971b). JOIDES drilling in late 1972 confirmed the Pliocene age of the Lau Basin. Neither basin has symmetrically linear magnetic anomalies as mid-oceans or the Scotia marginal sea do, although there are magnetic lineations in both (Chase, 1972; preliminary results of the 1971 Antipodes cruise).

(5) The Hunter Fracture Zone and its en echelon continuation west and north of Fiji is a complex trench-trench transform fault connecting the Tonga and New Hebrides subduction sites, and complicated by erratic spreading in the above-mentioned marginal seas. This interpretation is necessitated by general principles of plate tectonics (Wilson, 1965) and conforms to the gross structural trends emphasized by Hess and Maxwell (1949) and Carey (1963). The position of this fault is marked by a series of major seismic zones within which focal mechanism solutions indicate both left-lateral and thrust faulting (Sykes et al., 1969). In the west this fault separates the young, shallow Fiji Plateau from the older, deeper South Fiji Basin which has a thicker sediment cover. In the east it more dramatically separates the similarly young, shallow Lau Basin from the even older and deeper Pacific south of Samoa. This contact is seismically active (Sykes et al., 1969; Sclater et al., 1972; Barazangi and Isacks, 1972). Sparker profiles at about 175°W, 14.5°S show significant crumpling of sediments which contrasts sharply with the flat-lying sediments north and south of the fault (preliminary results of the 1971 Antipodes cruise).

Thus, modern Fiji is the emergent portion of a ridge of intermediate crustal thickness surrounded by young, currently-forming small ocean basins. This Lau-Colville Ridge is a "third arc" behind the Tonga-Kermadec Ridge, and the two appear to have separated from one another at a time equivalent to the age of the intervening (Lau) basin. Fiji therefore appears more closely tied to the history of Tonga than the New Hebrides. The maximum time Fiji could have existed in its present transitional environment is equivalent to the maximum age of the Fiji Plateau, Lau Basin, and Hunter Fracture Zone transform fault system.

2.3 A brief and rather romantic history of the Eastern Australian Plate

It is not possible in early 1972 to satisfactorally summarize the evolution of the eastern Australian plate. Many recent papers have elaborated and refined the older ideas of Suess (1904) and Glaessner (1950) that the existence of linear ridges and troughs east and roughly parallel to Australia is not geologically fortuitous. Packham (1972), Moberly (1972), Griffiths (1971a,b), Cullen (1970), van der Linden (1967), Rickard (in press) and others, have summarized the geology and bathymetry of this complicated region and attempted reconstructions which probably approximate an eventual synthesis. The following is a necessarily cursory statement of what is probably only personal belief in light of the many uncertainties and discrepancies, and the many subaerial and submarine areas of poorly-known geology and geochronology.

The Lord Howe Rise, Norfolk Ridge (including New Caledonia), and New Zealand Plateau appear to be continental fragments broken off from Gondwanaland. They have crustal thicknesses of 20 to 30 km and are separated by troughs with thinner, almost oceanic crustal profiles (Shor et al., 1971; Solomon and Biehler, 1969; Adams, 1962; Thomson and Evison, 1962). New Zealand, New Caledonia, and Papua, which define a bathymetrically discontinuous ridge subparallel to the Australian east coast, have very similar Paleozoic and Mesozoic lithologies, faunas, and histories. They share proximity to old sialic crust (Precambrian detrital zircons are known from New Zealand and New Caledonia: Aronson, 1968; Aronson and Tilton, 1971), and western sediment sources during the Permian to Jurassic (Brown et al., 1968, p.231; Avias, 1953), as well as sedimentological and volcanological evidence for subduction from the east during that period culminating in Early Cretaceous orogenies (Fleming, 1970; Dickinson, 1971a). Two

JOIDES drill cores from the Lord Howe Rise contain acid volcanics overlain by Mesozoic sediments.

Australasia (Australia-New Zealand-New Caledonia-Papua) seems to have faced continued subduction from the east, starting in at least the late Precambrian and continuing until the present day in the Tonga-Kermadec-North Island, New Zealand region, with a progressive eastward migration of subduction sites (Packham, 1972; Oversby, 1971; Brown et al., 1968). The opening of the Tasman Sea has been superimposed on this eastward migration of subductive foci. The Tasman is thought to be Cretaceous in age (see Vogt and Connolly, 1971; Hayes and Ringis, 1972; Griffiths, 1971a) and therefore similar in age to the South Pacific between New Zealand and Antarctica (oldest intervening magnetic anomaly is number 32; Pitman et al., 1968), in which case the Lord Howe Rise, Norfolk Ridge, and Campbell Plateau would have pivoted away from Australia and Antarctica simultaneously. Westward underthrusting of the Pacific beneath this moving leading edge of the Australian plate persisted into the Eocene at which time Australia separated from Antarctica (oldest intervening magnetic anomaly is number 17; LePichon and Heirtzler, 1968) and began its northward movement (Wellman et al., 1969). Spreading directions south of New Zealand apparently changed at this time (Christoffel and Ross, 1970).

Australia's drift brought with it the origins of Fiji. The outermost perimeter of the eastern Australian plate--Glaessner's (1950) "Outer Melanesian Arc" (OMA)--has an Eocene volcanic basement everywhere except the Solomon Islands (Tonga: Hoffmeister, 1932; Stearns, 1971; Fiji: Cole, 1960; New Hebrides: Coleman, 1969; New Britain: Macnab, 1970). The Pacific Province of the Solomon Islands has, and the Central Province may have a Cretaceous basement (Coleman, 1970), but at least the Pacific Province islands appear to have originated within the Pacific, to have been transported westward, and subsequently added to the OMA (Coleman, 1970; Hackman, 1972; Kroenke, 1970). Extensive Late Eocene to Early Miocene volcanism is preserved throughout the OMA (except the New Hebrides), includes much tholeiitic basalt and basaltic-andesite (Tonga: Ewart et al., 1972; Fiji: Gill, 1970 and chapter 4; Central Province, Solomon Islands: Hackman, 1972; Coleman, 1970; New Britain: Macnab, 1970), and apparently represents an island arc tholeiitic series. Amphiboles and plagioclase from Central Province, Solomon Islands basement rocks give

Eocene K-Ar ages (Richards et al., 1966; McDougall, pers. comm., 1971) the significance of which is ambiguous.

Gill and Gorton (Appendix 3) have therefore argued that a once-continuous island arc (New Hebrides-Fiji-Tonga Ridge) developed on the eastern perimeter of the Australian plate during the late Eocene to Early Miocene in response to yet another eastward migration of the subductive focus at which the Pacific plate continued to be consumed beneath the Australian. The South Fiji Basin (and an ancestral North Fiji Basin) presumably developed in the wake of this migration. JOIDES drilling found basic volcanics intruding Oligocene sediment in the South Fiji Basin.

Discovery of Eocene basalts east of the Tonga trench during late 1971 JOIDES drilling together with the inference of Ewart et al. (1972) that Eua may represent an ocean floor ophiolite sequence, casts doubt on whether the Tonga-Kermadec and Lau-Colville Ridges were ever one unit. If the non-volcanic outer ridge of Tonga is uplifted ocean floor yet the Lau Basin is no older than Pliocene, then Milsom (1970), Karig (1970), and Gorton and I are each partially but only partially correct. Modern Tongan volcanoes may occur on a horst of ocean floor (rather than older island arc crust) uplifted during inter-arc spreading, but this leaves unexplained the cause of mid-ocean floor Eocene volcanism and the apparent thickness of Tongan crust, and begs the question of how one distinguishes between mid-ocean and early island arc volcanism (see Jakes and Gill, Appendix 3).

Even if Tonga and Lau were once part of a single island arc underthrust by the Pacific plate, it is unclear whether this arc ever included the entire OMA or, if so, for how long. Ultramafic rocks and low-K submarine tholeiites appear to have been thrust from east to west over rocks of the Papuan and New Caledonian land masses (Avias, 1972; Lillie and Brothers, 1970; Davies, 1971; Coleman, 1971). The time of overthrusting is thought to be Late Eocene to Oligocene in Papua (Davies, 1971, p.10) although 50 to 55 m.y. K-Ar ages (i.e. Early Eocene: Evernden and Evernden, 1970) have been measured for tonalites thought to be related to this tectonism (Davies and Smith, 1971). Overthrusting in New Caledonia is thought to have occurred in the Oligocene to Early Miocene (K-Ar ages of 21 to 28 m.y.: Brothers, 1972; see also Lillie and Brothers, 1970, p.167; Coleman, 1967; Avias, 1972). These may, therefore, reflect two separate periods of

overthrusting. If the Oligocene to Lower Miocene age is correct for New Caledonia, this implies overthrusting of ocean floor crust behind the New Hebrides-Fiji-Tonga Ridge after it had become an active island arc, in which case it presumably would be inter-arc basin crust (of the South Fiji or former North Fiji Basins) that overthrust New Caledonia.

Did the present polarity of western Melanesian arcs, which contrasts strongly with Australasian Paleozoic and Mesozoic systems, persist throughout the life-span of Fiji? Northward underthrusting of the Australian beneath the Pacific plate seems unlikely to pre-date movement of Australia away from Antarctica (although the Late Cretaceous geology of Java suggests that the Australian-Antarctic plate underthrust the Asian plate then: Hamilton, 1970). As mentioned above, if the Eocene age of Papuan overthrusting is correct, initiation of Australian drift relative to Antarctica brought with it almost immediate collision with a presumably pre-existing subduction system north of Australia, resulting in overthrusting of oceanic mantle and crust onto the leading edge of the Australian plate. This could suggest an Andean-type pre-Eocene boundary with south-dipping subduction polarity between these plates, although most authors assume north-dipping subduction. The seismic division of Australia into four sub-plates (Cleary and Simpson, 1971) may stem from this collision and its resultant stress within the Australian plate.

Two indirect lines of evidence are consistent with (though not strong arguments for) continued west- and south-dipping subduction in the area west of Fiji throughout the period of Eocene to Early Miocene volcanism mentioned above. Mitchell and Warden (1971) infer westerly subduction beneath the New Hebrides in the Oligocene to Early Miocene from the occurrence of ultramafic rocks and amphibolites east of intermediate volcanic rocks thought (but not yet shown) to be similar in age. Macnab (1970, pp. 81-88) found two to three times more K_2O in intermediate plutonic rocks of southern New Britain than in rocks with similar SiO_2 contents further north, suggesting a south-dipping Benioff Zone. These rocks appear to be Oligocene to Middle Miocene in age (Page in Macnab, 1970, p.59, and pers. comm., 1972).

Hamilton (1970), Dewey and Bird (1970), and Johnson and Molnar (1972) proposed that, at some later time, New Guinea also over-ran a westward extension of the New Britain Trench. Page (1971) suggests

that this collision produced the New Guinea Mobile Belt, the plutons of which he found to be 12-15 m.y. old. JOIDES drilling in the Coral Sea found Middle Miocene turbidites, consistent with uplift in New Guinea at this time. If the pre-Miocene history of North New Guinea includes evidence of an island arc and if that island arc was continuous with New Britain and lay above underthrusting of the leading edge of the Australian plate, then the Eocene to Middle Miocene subduction polarity of western Melanesian island arcs probably was then as it is now. Otherwise, the opposite polarity may have persisted until the Middle Miocene. It is, therefore, uncertain whether the northeast-dipping subduction environment, now found west of Fiji, existed early in Fiji's history.

I shall return to these issues in chapter eight.

Chapter 3. The Geochemical Context: spatial variations in the composition of island arc magmas and rocks

3.1 Introduction

For half a century, Japanese and Dutch geologists have been saying that the composition of magmas, and especially the relative and absolute alkali abundances, varied across the strike of island arcs. This realization gained most influence through the papers of Sugimura (1961, 1968, 1972), Kuno (1959, 1966), and Dickinson and Hatherton (1968). Such regularity provides one of the few windows through the general opacity of analytical data in the security of which various possibilities for island arc magma genesis reside unconstrained. Furthermore, such regularity leaves clues within island arc volcanic rocks with which to ascertain at some later time the tectonic polarity of their subduction environment (Dickinson 1970). This regularity is the basis for my argument that Fiji developed as a normal arc and only recently acquired its enigmatic character.

World-wide and/or circum-Pacific summaries of island arc magma compositions by Kuno (1966), Dickinson (1968), and Gorshkov (1970) have indisputably demonstrated transverse variations in alkali, principally K_2O, contents. Sugimura (1968, 1972) has used multiple correlation coefficients to assess variations in major element oxides and normative minerals. I have tried to isolate and emphasize variations in major, minor, and trace elements and isotope compositions--see all papers in Appendix 3. Table 3.1 and Figures 3.1 to 3.3 summarize, in an inevitably subjective manner, variations which have been documented in modern island arcs. The data are of uneven quality and from numerous articles, and rarely are the same rocks analyzed for all elements or even well-described petrographically or geographically. This chapter can only attempt a preliminary examination of these variations. An additional and similar summary is given by Jakes and White (1972).

Some conventions used in the table, figures and text which follow merit comment. To facilitate geographical references, discussion is in terms of distance from the trench rather than depth to the seismic zone even though K_2O correlates better with the latter than the former (Hatherton and Dickinson, 1969). The discussion also emphasizes variations within rather than between arcs and avoids absolute values of distance or depth. Most of the geochemical parameters used vary

Table 3.1. Geochemical variations in volcanic rocks from modern island arcs, related to distance from the adjacent trench.

	JAPAN: northern Honshu			KURILES		KAMCHATKA			ALEUTIANS	
	Tholeiitic Series	High-Alumina Series	Alkalic Series	Main Zone	Western Zone	East Coast	Central Depression	Sredinnyi Ridge	Main Zone	Bogoslof
Increasing Distance from Trench	→			→		→			→	
SiO_2 range	48-68	46-70	49-62	48-75	51-59	51-56	52-65	46-73	49-76	61.
TiO_2	1.2-0.5	1.0-0.5	1.1-0.4	1.1-0.4	1.3-0.7	1.0-0.6	1.0-0.4	1.3-0.4	1.2-0.2	0.5
Max. FeO) + Fe_2O_3)	13.	9.5	10.5	11-12	9-10	11-12	9.	9.	12.	5.
K_{60}	0.7-1.0	1.2-2.2	3.3	1.1	2.	0.7-1.3	1.5-2.	2-3	1.5-2.	3.
P_2O_5	0.1	0.2-.3	0.2	0.1-.2	0.2-.4	0.1-.2	0.2-.4	-	0.1-.3	0.3
Na_2O/K_2O	5-10	3-4	1.2	3-4	1.2-2.	3-4	1.2-2.	0.4-2.	2-4	1.3
Rb	0.3-5.	30-50.	80.	-	-	-	-	-	-	-
Sr	180-300.	700.	500-800.	-	-	-	-	-	-	-
Ba	70-120.	250-350.	-	-	-	-	-	-	-	-
Pb	1.5-6.	2-20.	2-5.	-	-	-	-	-	-	-
Th	0.2	1.3-12.	2-6.	-	-	-	-	-	-	-
U	0.1	0.5-3.	0.5-1.5	-	-	-	-	-	-	-
La	1.5-6.	8-15.	-	-	-	-	-	-	-	-
Yb	1.6-2.2	0.5-3.	-	-	-	-	-	-	-	-
Sr87/86	0.7041-3.	0.7032-6.	0.7026-31	-	-	-	-	-	-	-
K/Rb	850-1800.	400-600.	200.	-	-	-	-	-	-	-
La/Yb	1-2.5	6-8.	-	-	-	-	-	-	-	-
Th/U	2-3.	2-4.	4.	-	-	-	-	-	-	-
References	(1-12)			(13)		(13-15)			(16-18)	

Table 3.1. Continued

	NEW BRITAIN		JAVA			MARIANA-IZU		RYUKUS-WEST JAPAN	
	North Coast	Talasea	Soropati	Merapi	Ungaran	Main Zone	Volcano Islands	Main Zone	Unzen-Aso
Increasing Distance from Trench	→		→			→		→	
SiO_2 range	50-75.	51-75.	54.	55.	56.	51-76.	58-60.	51-72.	52-72.
TiO_2	0.6-0.4	1.0-0.3	0.9	0.7	0.7	1.1-0.4	1.1-0.7	1.2-0.4	1.2-0.4
Max. FeO) + Fe_2O_3)	10.	11.	9.4	7.7	7.7	13.	7.	10.	10.
K_{60}	0.8	1.5	1.6	2.1	2.9	0.8	4.0	1.0-1.5	2.0
P_2O_5	0.1-.2	0.2	0.2	0.3	0.3	0.1-.2	0.4-.5	0.1-.2	0.1-.3
Na_2O/K_2O	3-6.	2-4.	2.0	1.8	1.2	3-10.	1.5-2.	2-4.	1.2-2.
Rb	-	5-60.	35.	49.	83.	-	-	-	-
Sr	250-400.	200-500.	470.	570.	485.	-	-	-	-
Ba	100-200.	200-600.	450.	540.	570.	-	>1000.	-	-
Pb	-	-	13.	18.	15.	-	-	-	-
Th	0.1-.5	-	6.0	8.2	14.	-	-	-	-
U	0.1-.5	-	1.3	1.8	2.3	-	-	-	-
La	-	-	-	-	-	-	-	-	-
Yb	-	-	-	-	-	-	-	-	-
Sr87/86	0.7035	0.7035	-	-	-	-	-	-	-
K/Rb	-	500-100.	375.	360.	290.	-	-	-	-
La/Yb	1-2.	-	-	-	-	-	-	-	-
Th/U	-	-	4.6	4.5	6.0	-	-	-	-
References	(19-22)		(22)			(2,23)		(2,13)	

References:

1. Minato et al. (1962)
2. Kuno (1962)
3. Ono (1962)
4. Sugimura (1961)
5. Hedge & Knight (1969)
6. Tatsumoto & Knight (1969)
7. Masuda (1966, 1967)
8. Philpotts et al. (1971)
9. Schnetzler & Philpotts (1970)
10. Philpotts & Schnetzler (1970)
11. Taylor & White (1966)
12. Nagasawa et al. (1971)
13. Gorshkov (1970)
14. Ehrlich (1968)
15. Vlodavetz & Piip (1959)
16. U.S.G.S. Bull. 1028 A-T.
17. Byers (1961)
18. Coats (1962)
19. Lowder & Carmichael (1970)
20. Peterman et al. (1970a)
21. A. Ewart, W. Johnson, R.Page (pers. comm.)
22. Gill, new data
23. Schmidt (1957)

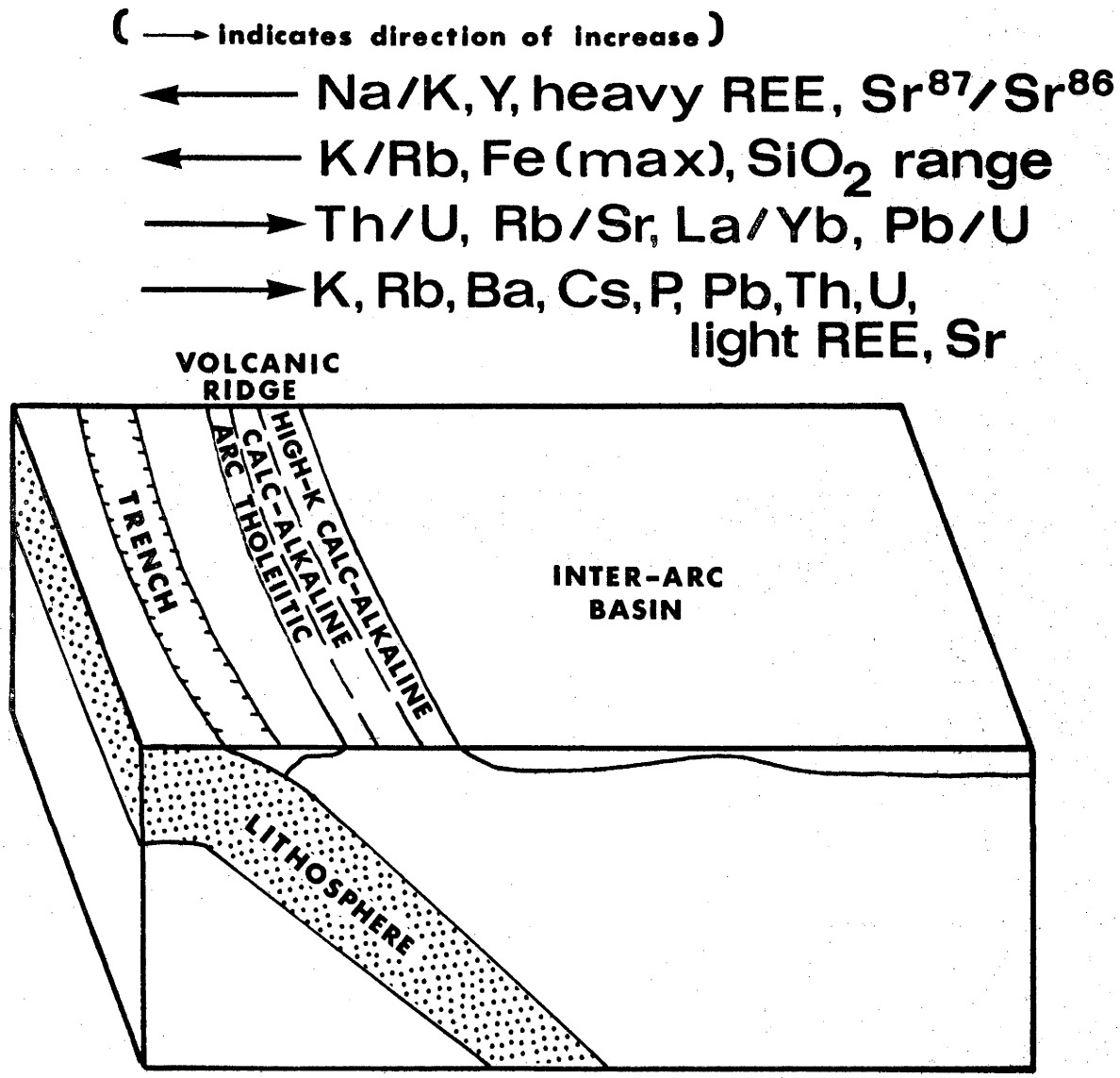

Figure 3.1 Spatial variations in the geochemistry of volcanic rocks in island arcs. Revision of Figure 1 in Gill and Gorton (Appendix 3). Data are taken from Table 3.1; nomenclature from section 1.3.

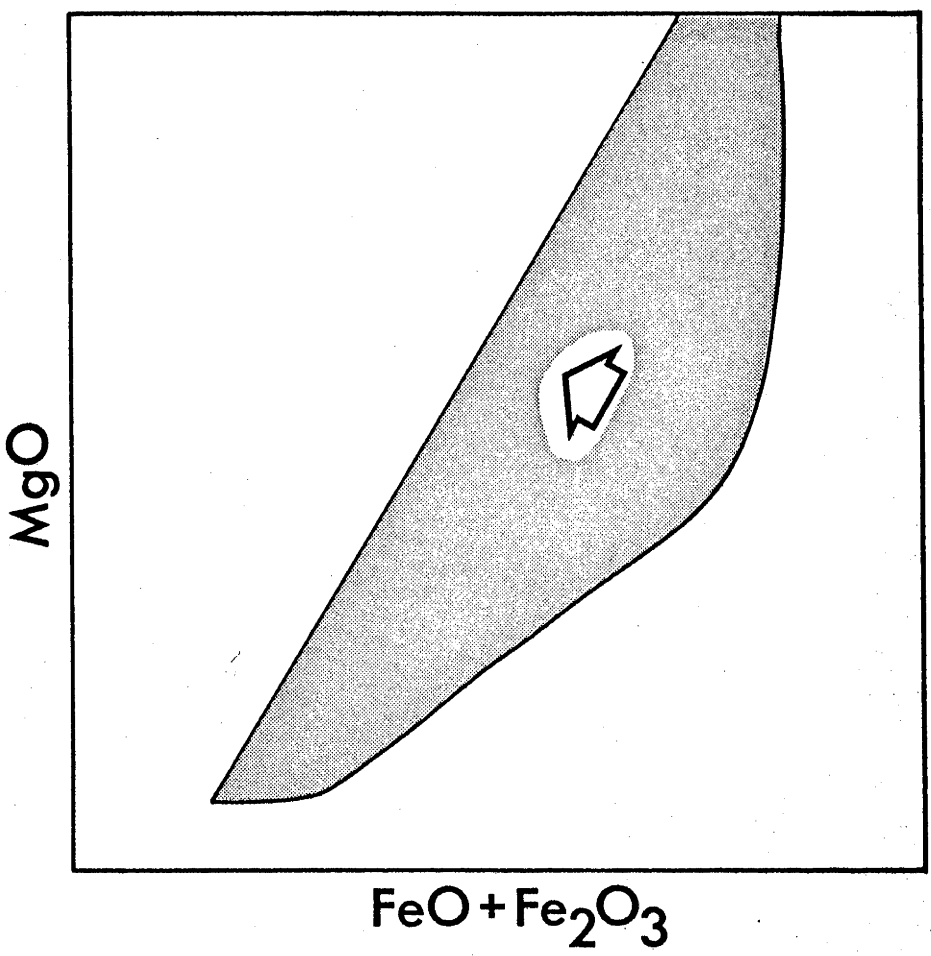

Figure 3.2 Most common variation of iron-enrichment patterns in island arcs. Arrow indicates distance from the trench. Calc-alkaline rocks define relatively straight lines; in more tholeiitic series Fe contents initially increase while Mg contents decrease. The relationship to distance from the trench is common but not universal.

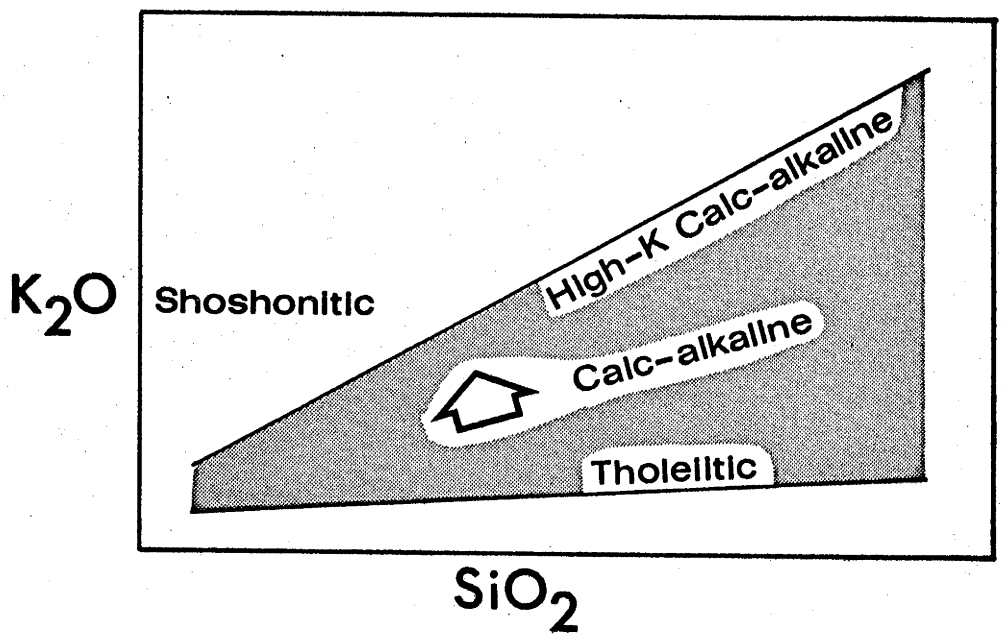

Figure 3.3 Variations of K_2O and SiO_2 in island arcs. After Dickinson (1968a), Gill (1970, Figure 8), Jakes and Gill (1970, Figure 2), and Smith (1970, Figure 12). Arrow indicates distance from the trench.

with SiO_2 and/or MgO contents, as recognized by Dickinson and Hatherton but not by Hart et al. (1970). Strict meaningfulness would be possible only when equivalent samples are compared, but this is usually precluded by lack of data. Moreover, "average" SiO_2 and MgO contents apparently change with distance from the trench (Sugimura 1968; Jakes and White, 1972) so that samples with equivalent Si and Mg may not be most representative of their respective arc positions. I therefore did not attempt quantitative rigor in Table 3.1 and tried only to summarize what appeared most representative of the areas described.

3.2 Spatial variations in the geochemistry of modern island arcs

A. Japan

Sugimura and Kuno dealt mostly with Japan so it has become paradigmatic for island arc magma variations. This is a mixed blessing because Japan is a complicated area involving at least two different arc systems--the East (EJVB) and West (WJVB) Japan Volcanic Belts (Sugimura, 1968, Figure 1). The former separates the Asian and Pacific plates; the latter separates Asia from the Philippine Sea plate(s). These converge near the Fossa Magna of southeastern Honshu, are separated by a marginal sea, and represent arms of a complex triple junction (Fitch and Scholz, 1971). In Kuno's (1966) widely accepted subdivision of island arcs into zones of primary tholeiitic, high-alumina, and alkali basalt volcanism, the differences between these arc systems are muted. The WJVB has no tholeiitic, some high-alumina, and many alkalic rocks. The EJVB has many tholeiitic, less high-alumina, and few alkalic rocks (Yagi et al., 1963; Kuno, 1966, Figure 1; Sugimura, 1968, Figure 10). More significantly, the alkalic rocks of the two arcs are quite dissimilar, although they have remained grouped together ever since Tomita (1935) defined the "Circum-Japan Sea Cenozoic alkali province". Those of the EJVB have $TiO_2 \leq 1\%$, K/Na = 0.8, and $^{87}Sr/^{86}Sr < 0.703$, all typical of alkalic (high-K calcalkaline and shoshonitic) rocks of island arcs; whereas those of the WJVB have TiO_2 = 2 to 4%, K/Na = 0.5, and $^{87}Sr/^{86}Sr > 0.704$, typical of intra-oceanic alkali basalts. The only island arc analogues of WJVB alkali basalts (Pribilof, Aleutians; Fiji, chapter 7) are also associated with a marginal sea environment and may be more related to inter-arc rifting than primary underthrusting.

Even within each Japanese arc system, however, Sugimura (1961, 1968) and Gorshkov (1970) have vindicated Kuno's emphasis on the

importance of potassium variations, but attempts to evaluate other geochemical parameters meet two difficulties, apart from the paucity of data. The first was Kuno's unwavering identification and use of groundmass pyroxene compositions to classify rock series which has baffled or outskilled most petrologists. His distinction between pigeonitic and hypersthenic series has geochemical overtones (chapter 11) but no consistent spatial significance. The former is largely coincident with the tholeiitic province and characterized as much by low K_2O as low Al_2O_3 (Kuno 1960, Figure 5). The hypersthenic rock series is not necessarily coincident with the high-alumina province, as might be inferred by conjoining Kuno's 1959 and 1960 papers, and lacks spatial significance. (In the Hakone region, for example, rocks of the two series are interbedded [Kuno, 1950]). Thus, the important spatial zonation in northern Japan is tholeiitic--high-alumina--shoshonitic. Rocks identified only with the pigeonitic series could occur in either of the first two zones; those from the hypersthenic series could have come from anywhere. (To Kuno, each of his primary basalt types could, upon fractionation, produce either its own distinctive differentiates or produce predominantly intermediate volcanics lacking iron-enrichment but containing groundmass hypersthene [Kuno, 1968a,b]. The latter were calc-alkaline rocks to Kuno and he used nouns like andesite and dacite independently of rock series as I illustrated in Figure 1.2).

The second handicap is that most data pertain to rocks from the Fuji-Hakone-Izu region which, although well-studied by Kuno and important to the history and culture of Japan, is a poor type-locality for variations in island arc magma compositions. It is within the tectonically complex triple junction mentioned above and, perhaps consequently, there is scant spatial separation between the tholeiitic and high-alumina provinces. Moreover, analyzed rocks from this region are often identified only with the pigeonitic or hypersthenic rock series without regard to spatial relationships.

The available data nonetheless define a consistent pattern (Table 3.1). Hedge and Knight (1969) report that the isotopic composition of Pb also becomes less radiogenic further from the trench in this region. Tatsumoto and Knight (1969) reach the same conclusion but only by combining samples from the East and West Japan Volcanic Belts.

B. The Kuriles

Gorshkov (1970, pp. 239-271) has shown the ten volcanoes of his Western Zone to be richer in K and P, less Fe-enriched, and to have a contracted SiO_2 range, relative to those of his main zones.

A slight westward increase in U is apparent from the data of Leonova and Udaltzova (1970). It is, however, obscured by data scatter and by disagreement with data of Kravchenko and Semenov (1970). The former authors report an average of 0.25 ppm U, 0.35 ppm Th, 10 ppm Rb, and K/Rb = 550 for basaltic-andesites and andesites of the main Kurile zone.

C. Kamchatka

Similar variations in K, P, and Fe have been discussed by or may be seen in the data of Ehrlich (1968), Vlodavetz and Piip (1959), and Gorshkov (1970, p. 281). There are few published trace element data for rocks from Kamchatka and even those lack geographic control.

D. The Aleutians

Of the approximately 40 active Aleutian volcanic islands (excluding Pribilof), only Bogoslof and Amak lie behind the main volcanic front (Byers, 1959, Figure 50). Two major element analyses of rocks from Bogoslof have been published; none from Amak. There appears to be considerable diversity within the main zone itself but less than the difference between it and Bogoslof.

E. The Americas

I excluded the Americas from Table 3.1 due to the tectonic complexity of their Pacific boundary and the greater likelihood of crustal anatexis or modification, at least in Central and South America (see Jakes and White, 1972). Qualitatively similar trends are nevertheless present in data for their Quaternary volcanics. K and Rb increase west to east in the Southern Cascades (Smith and Carmichael, 1968). K, Rb, Ba, Ti and the Si mode increase west to east in central Mexico whereas K/Rb ratios decrease (Gunn and Mooser, 1970). K, Rb, Th and Y increase west to east across northern Chile whereas Sr and Ni decrease (Rhodes and Ridley, 1971).

F. The Antilles and Scotia arcs

No volcanoes lie behind the volcanic front of either of these arcs except for Lestov, slightly to the rear of the northern Scotia trend. No analyses of its rocks have been published but Baker

describes them as "two-pyroxene andesites and rather different from the lavas found along the main arc" (written comm., 1971).

G. New Zealand

I excluded New Zealand data from Table 3.1 on the same grounds as for the Americas. Ewart and Stipp (1968), Black (1970), and others have demonstrated the role of crustal involvement in intermediate and acid magma genesis there. Nevertheless, Cole (1967) and Stipp (1968) have shown a general westward increase in K, Rb, Th contents and Rb/Sr ratios, and decrease in $^{87}Sr/^{86}Sr$ ratios of basalts.

H. Tonga-Kermadec

All volcanoes lie at this volcanic front. Fein (1971) claims a westward increase in K, Rb, and Ba and decrease in V but, presumably, only by including analyses of rocks from Eua which is older or Niua Fo'ou or Fiji which are related to rifting between Tonga and the New Hebrides (Gill and Gorton, Appendix 3; Sclater et al. 1972).

I. The New Hebrides and Solomon Islands

These volcanoes, too, define one boringly undeviant volcanic front. Mallick (1972) suggested a west to east decrease in K in the Banks Islands (i.e. opposite to the K variations in this summary), but Gill and Gorton (Appendix 3) show that this variation may instead reflect the current rupturing of the New Hebrides Ridge.

J. New Britain and New Guinea

West of Rabaul, New Britain is a fairly straightforward and increasingly well-studied arc. Johnson (1970) demonstrates northward increase in K_2O. The K, U, and Th data in Table 3.1 for volcanics of the New Britain north coast are a summary of my own analyses; other data are the unpublished results of A. Ewart, W. Johnson, and R. Page (pers. comms., 1971).

Jakes and White (1969) reported spatial variations in K_2O and K/Na for New Guinea which are consistent with the trends emerging from this summary but I have omitted them due to the tectonic complexity of that area (see Denham, 1969; Johnson et al., 1971).

K. Indonesia

Although Indonesia was the second area from which spatial variations were reported in island arc magma compositions (Rittman, 1953), little analytical work followed the compilation by Neuman van Padang (1951). Hatherton and Dickinson (1969) summarized that

compilation and demonstrated the correlation between K and distance from the trench in Java and Sumatra. No trace element or isotope data have been published for Indonesian volcanics. The entries in Table 3.1 are from analyses of single rocks taken from three volcanoes at different distances from the Java Trench and sent to S.R. Taylor by V. Sen in 1969. Although a tholeiitic series appears non-existent and alkali contents are generally high, transverse variations are similar to those observed elsewhere.

L. Mariana-Bonin-Izu islands

This and the following section complete the circum-Pacific summary by returning to Japan. Most of those islands which form the south-eastern extension of the East Japan Volcanic Belt define a linear volcanic front, but the so-called "Volcano Islands" lie about 50 km behind this front, coincide with shoaling of the trench, and are enriched in K_2O.

M. Ryukus-West Japan Volcanic Belt

Again there are some volcanoes behind the front but scant major and no trace element data with which to assess variations other than in K_2O contents.

N. Aegean

Nicholls (1971) demonstrated increasing K and P and decreasing Na/K and Fe as the distance between trench and volcano increases.

3.3 Varieties of alkali-rich rocks in island arcs

Most surveys of K variations by Kuno and Sugimura, Dickinson and Hatherton, or Jakes, White and Gill have loosely alluded to alkali basalts, shoshonites, or high-K calc-alkaline rocks in island arc hinterlands. Simultaneously, Karig and others have emphasized the occurrence of inter-arc rifting and its associated volcanism in those hinterlands. This volcanism appears to be largely tholeiitic (Sclater et al., 1972) rather than alkalic as proposed by Packham and Falvey (1971), but high-Ti alkali basalts are spatially conjoined with rifting sites in Japan and the Aleutians. High-Ti alkali basalts also occur in Fiji and I have argued that they and the Fijian shoshonites post-date subduction and, instead, reflect inter-arc rifting (Chapters 8 and 12 and Gill and Gorton, Appendix 3). A similar argument may also apply to the alkali and transitional basalts of the New Hebrides. Thus the various high-K rocks in island arcs may have quite different

causal explanations and the mafic, high-Ti and intermediate, low-Ti rocks certainly do.

3.4 Stratigraphic variations in the geochemistry of island arc volcanic rocks

Earlier (Gill, 1970; pp. 194-6) I summarized instances of and references to situations where the geochemistry of volcanic rocks may vary with time just as with distance from a trench, younger rocks being similar to those further from a trench. In Gill and Gorton (Appendix 3) I argue that the cause of this space-time symmetry is migration of trenches toward their associated spreading-center with time. Such migration is implicit in the idea of continental accretion, has been inferred from age relationships of orogenic belts in southeast and east Asia (Hamilton, 1970; Dickinson, 1972) and age relationships of western Pacific island arcs (Karig 1971a,b), and should be expected during subduction characterized by enlargement of the hanging wall due to sediment accretion and by dense lithosphere overlying less dense mantle (see Moberly, 1972; Oxburgh and Turcotte 1971, p. 1326; Elsasser 1971; and Ringwood, 1972). If trench migration occurs, the distance between some position on an island arc ridge and either the trench or Benioff Zone may increase with time, resulting in the stratigraphic variation commonly observed. If trench migration is discontinuous it may be accompanied by uplift of the ridge and formation of stratigraphic unconformities. The effects of trench migration are sketched in Figure 3.4.

Note two consequences of this explanation. First, there is nothing *a priori* about temporal geochemical variations; the apparent consistency is due only to consistent migration of orogenic foci. In areas such as Honshu, Japan where arc tectonics has been complex, stratigraphic regularities in geochemical variations should not be expected and are not observed (Minato *et al.* 1962). Second, there will be a strong *primary* geochemical zonation of island arc crust with upward enrichment of alkali and other large cations, ^{87}Sr, and radiogenic Pb, as has also been noted by Jakes and White (1971). Subsequent upward concentration of these elements due to partial melting or metamorphism (Heier, 1965; Lambert and Heier, 1967; Lachenbruch, 1968) will enhance not initiate this zonation.

3.5 Summary and causes of the variations

The data from Table 3.1 have been summarized in Figures 3.1 to 3.3. Elucidation of the reasons for them requires explanation of

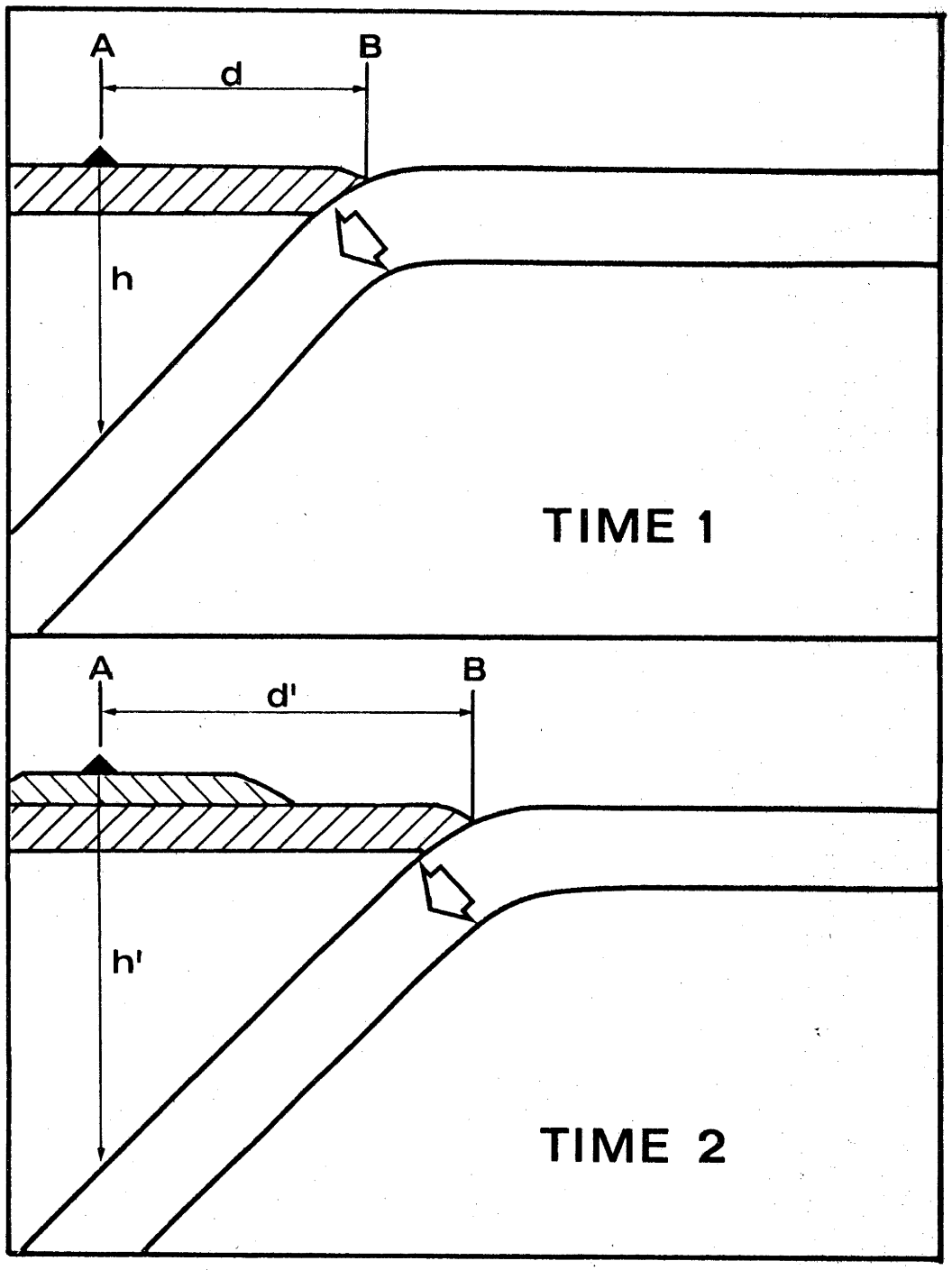

Figure 3.4 Effects of trench migration on temporal variations in island arc geochemistry. Crust at Time 1 might be due to island arc tholeiitic series volcanism corresponding to some distance h from the Benioff Zone and d from the trench. If the trench migrates toward its associated spreading center, at some later Time 2 the Benioff Zone will be deeper beneath volcano position A (h'>h) which will also be further from the migrating trench (d'>d). Younger crust may be due to calc-alkaline volcanism. In this way stratigraphic variations may come to reflect the spatial variations in geochemistry shown in Figures 3.1 to 3.3.

island arc magma genesis itself and will be considered more in Chapters 9 to 11.

Note, however, that although many of the elements enriched in magmas erupted further from the trench are those also enriched in sialic crust, their variation in island arc rocks is relatively uniform and irrespective of crustal thickness, age, or structure. These variations are therefore not likely to result from increasing crustal contamination with distance from the trench, as Gorshkov has repeatedly emphasized about alkali variations. Neither can they be explained by simply invoking greater degrees of "wall-rock reaction" as the amount of mantle through which ascending magma must pass increases (T. Green 1972). These issues are discussed more fully in sections 9.2, 9.3 and 10.8.

Chapter 4. Geology and Geochemistry of Viti Levu

4.1 Introduction

Viti Levu is the largest Fiji island (10,000 km^2) and contains a majority of the new nation's population. Recorded exploration of its interior began in 1856 and initial geologic mapping at 1:50,000 scale was finished during preparation of this thesis. The geology of Viti Levu was summarized by Rodda (1967) and there are petrographic descriptions of its igneous rocks in many FGSD publications. Nothing has been added since my earlier compilation from these sources (Gill, 1970) so I will not repeat myself here.

The division of Viti Levu's volcanic history into three temporally and geochemically distinct periods which I adapted previously (Gill, 1970) from Hirst (1965, Table 2), appears sound and will be followed again. The geochemistry and age of plutonic rocks will be considered separately in section 4.5. Locations of the 101 Viti Levu samples analyzed are indicated in Figure 4.1 and listed together with petrographic descriptions in Appendix 2.

4.2 First period volcanism: Eocene to early Middle Miocene

A. Age, environment, and thickness

Rocks of the Wainimala, Singatoka, and Savura Groups record events of this period. They are largely volcanic or volcaniclastic and many were submarine (Band, 1968; Hirst, 1965; Houtz, 1960, 1963), although no unequivocally abyssal rocks are known. Even the oldest known fauna probably lived in water shallower than 70 m (Cole, 1960). Thus, although this probably was a period during which Fiji rose as a submarine ridge, its earliest memories are either not preserved, exposed, or recognized.

The volume and maximum age of these rocks is uncertain. As outlined in section 2.3, basements throughout the Outer Melanesian Arc (excluding the Solomon Islands Pacific Province) expose Eocene volcanic, volcaniclastic, and metamorphic rocks; no older fauna or radiometric ages have been reported. Crustal thickening due to recent island arc volcanism can be calculated as approximately 0.05 to 0.2 km/m.y. from data for the Kuriles and the East Japan Volcanic Belt (Markhinin, 1968, p. 419; Sugimura, 1968, Figure 10 and Table 4). Continuous production of new crust (i.e. without remelting pre-existing crust) over 50 m.y. at either rate would not yield enough material to produce the gravity anomalies observed over Viti Levu

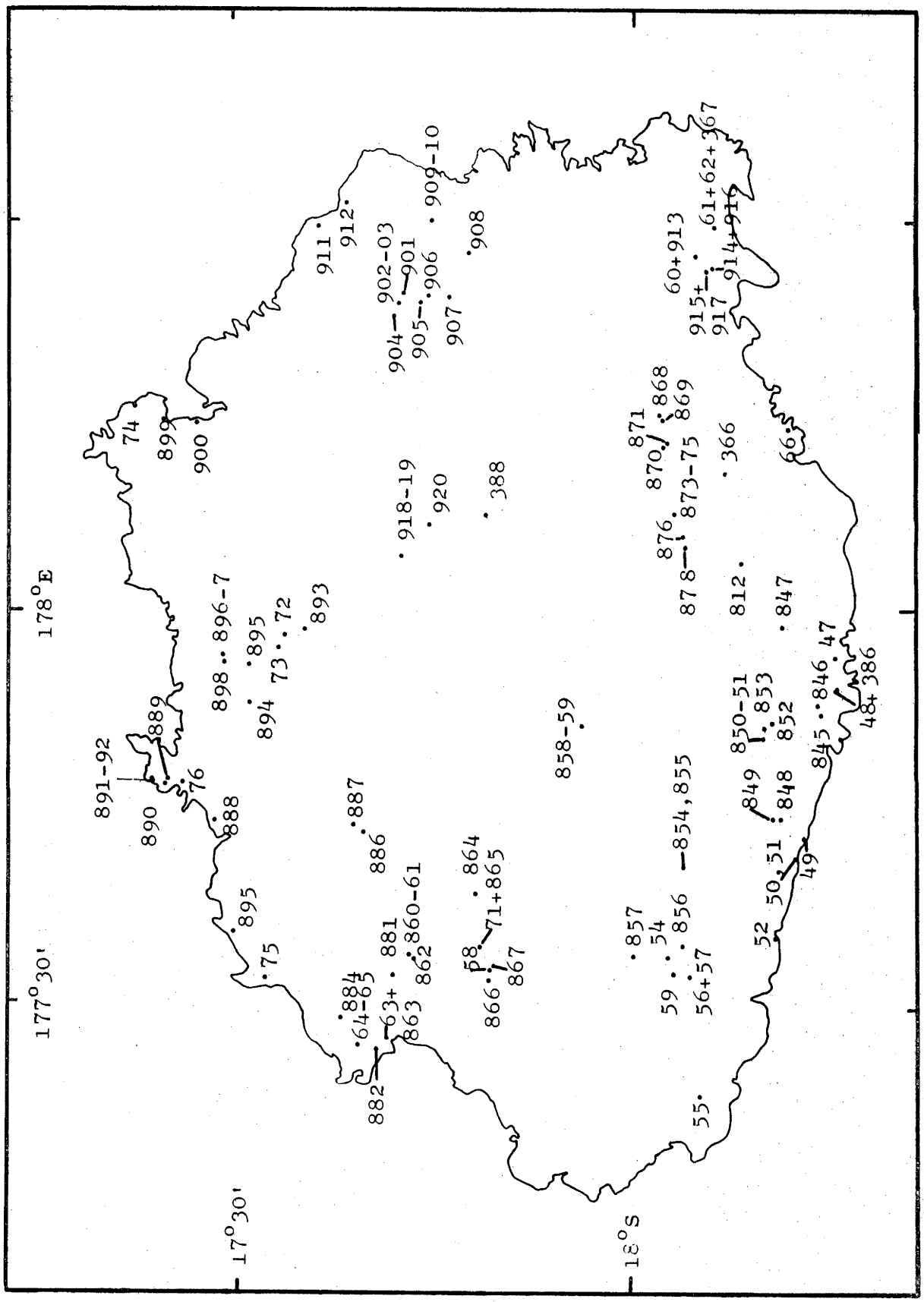

Figure 4.1 Locations of Viti Levu samples

(Robertson, 1967) if the mean density of such crust is the 2.7 g/cm^3 measured for Wainimala rocks. If Viti Levu is less than 50 m.y. old, then (a) its early rate of volcanism must have been very great (as suggested for Neogene volcanism in Japan by Sugimura et al., 1963), (b) its crust includes much residual material left below by magmatic crystal fractionation (as argued by Kuno, 1968a, pp. 168-171), (c) older crust moved eastward along with migration of the subduction zone (as Eua and Saipan appear to have: Karig, 1971a,b) and participates in the basement of Fiji, or (d) the crustal column contains oceanic sediments accreted during plate subduction.

It is not known whether volcanism and deposition were continuous from Late Eocene to early Middle Miocene or were sporadic. Fauna of the older age have been found only in southwestern Viti Levu; of the younger age, only in the east. Submarine volcanics of similar composition are associated with both faunas.

Many first period rocks have been metamorphosed to zeolite and green-schist facies assemblages of which the best descriptions remain those of Crook (1963). Metamorphic effects are less near the top of the stratigraphic section (i.e. eastern and northeastern areas of outcrop). These rocks have also been deformed to varying degrees (see Dickinson, 1967, p. 548) during what Band (1968) called the Tholo Orogeny. Tectonism was not severe and may have accompanied or just preceded Middle Miocene uplift and development of the erosional unconformity which everywhere truncates the record of this period.

B. Petrography

Although most samples have experienced some post-magmatic mineral reconstitution, a continuum from very fresh to almost completely albite+chlorite rocks were studied so that a few generalizations about their initial and altered characteristics are possible. Phenocrysts are usually <20 modal per cent and show that most lavas began to crystallize under conditions in which plagioclase and clinopyroxene were their liquidus or near-liquidus phases. Olivine, orthopyroxene, and hornblende are rare; small oxide phenocrysts are common but rarely exceed two or three modal per cent. The groundmass usually is fine-grained, consisting of flow-oriented plagioclase laths with interstitial pyroxene and oxide grains.

Many rocks are altered but in very few have feldspars lost their zoning and twinning. Recrystallization is often confined to areas of

devitrified glass where nucleation is easiest, both in the groundmass and as inclusions in plagioclase.

C. Geochemistry

Most samples for which analyses are listed in Tables 4.1 and 4.2 were collected from areas mapped as containing rocks belonging to formations of the Wainimala, Singatoka, or Savura Groups. Because it was not always possible to obtain unweathered rocks in situ, some samples are stream cobbles with inherent stratigraphic uncertainty. Samples 905, 866 and 855 are dense with fine-grained diabasic textures and may be from dikes or sills; the former two are stream cobbles and the latter looked like a flow in outcrop. Samples 860 and 862 are from outcrops mapped as belonging to the younger Nandi Sedimentary Group (Bartholomew, 1960; Rodda and Band, 1966). I include them here because the mapping is suspect and Wainimala Group rocks are expected in that area (Rodda, pers. comm., 1971), and because of their composition. Seven other major element analyses of first period rocks are compiled by Rodda (1969) of which three are tuffs and two dikes.

The 43 available analyses define a suite with a broad SiO_2 range and gradually varying composition, the characteristics of which are displayed in Figures 4.2 and 4.3. (Harker diagrams have no genetic significance and only the Ti, P, and K variances are likely to be statistically meaningful [Chayes, 1964a]. The "trends" are merely graphical depictions of how similar in composition magmas of a given silica content remained throughout this period of volcanism.) Note especially the low K_2O and P_2O_5, relatively high Fe_2O_3*, the low TiO_2 contents and their narrow range, and the variable but high Al_2O_3. $Mg/Mg+Fe^{+2}$ ratios are generally <0.60 in the basalts but slightly higher than in other rock types where they average between 0.48 and 0.52 and are statistically indistinguishable from one another. Na_2O/K_2O ratios are variable and generally >4.

The suite is the least alkali-rich of any in Fiji. K, Rb, and Ba are variable but low as are Sr and Pb. Th and U are rarely detectable (i.e. <0.1 ppm); Th/U ratios when measureable are 1 to 2. Average heat production from such rocks would be <0.5 μcal/g yr.

Transition metal contents are also low. Ni is low even relative to Mg; Mg/Ni ratios average 2400 in the basalts and all but two are >1000. Ni/Co ratios are low and V/Ni high throughout the suite, and V contents drop considerably though erratically.

TABLE4.1. FIRST PERIOD VITI LEVU VOLCANICS.

DATA	862	61	62	905	367	853	907	855	920	860	850	49	878	851	904	54
SIO2	47.37	47.68	47.98	48.57	49.20	49.37	49.54	49.59	49.72	50.11	51.91	52.76	54.47	54.77	55.60	56.74
TIO2	1.20	1.13	1.18	1.57	1.07	0.80	1.09	0.99	1.07	1.26	0.79	0.74	1.04	0.73	0.79	0.59
AL2O3	17.44	18.86	19.41	16.86	18.51	18.66	15.18	19.28	18.37	18.91	20.03	18.63	17.67	18.37	18.55	18.63
FE2O3	10.83	7.84	9.39	13.14	11.12	11.82	11.56	9.99	11.59	10.24	10.05	3.89	10.03	10.42	8.27	4.32
FEO	0.0	3.86	2.82	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	6.08	0.0	0.0	0.0	3.95
MNO	0.19	0.24	0.22	0.24	0.18	0.20	0.17	0.17	0.19	0.19	0.18	0.20	0.18	0.17	0.17	0.22
MGO	4.80	6.23	5.35	7.58	5.52	5.13	6.78	4.97	4.64	4.88	3.40	4.92	3.40	3.43	4.04	3.67
CAO	12.90	10.06	9.59	7.23	10.37	11.20	10.51	11.52	11.50	9.84	8.20	7.36	9.24	7.88	8.68	8.29
NA2O	4.01	2.82	2.95	4.21	2.77	2.63	2.87	3.00	2.31	3.17	4.62	4.11	3.22	3.78	3.10	2.72
K2O	1.04	0.97	0.79	0.40	0.96	0.12	1.90	0.36	0.44	1.21	0.76	1.24	0.61	0.39	0.63	0.68
P2O5	0.23	0.30	0.32	0.19	0.31	0.07	0.40	0.16	0.16	0.19	0.08	0.06	0.14	0.06	0.16	0.19
(LOI)	7.67	4.32	4.67	4.01	1.50	3.02	1.93	3.94	0.50	2.99	2.63	3.81	0.0	2.31	0.69	3.14
(TOTAL)	99.41	99.75	98.87	99.39	99.71	99.59	99.30	99.37	99.01	99.29	99.24	99.47	98.31	99.52	99.59	99.66
NA2O/K2O	3.9	2.9	3.8	10.6	2.9	21.2	1.5	8.4	5.3	2.6	6.1	3.3	5.3	9.7	5.0	4.0
MG/MG+FE2	0.52	0.56	0.51	0.58	0.55	0.51	0.59	0.55	0.49	0.54	0.45	0.53	0.45	0.44	0.54	0.51
%K	0.86	0.81	0.65	0.33	0.79	0.10	1.58	0.30	0.36	1.01	0.63	1.03	0.51	0.32	0.52	0.57
RB	12.	8.3	5.0	33.	12.	1.3	33.	4.4	5.2	7.9	6.1	6.1	6.5	3.6	8.9	9.5
SR	332.	619.	522.	459.	583.	164.	778.	319.	283.	215.	156.	114.	311.	138.	357.	373.
BA	77.	169.	190.	145.	137.	43.	375.	42.	57.	83.	97.	130.	96.	94.	133.	191.
PB	0.	3.	3.	4.	2.	0.	4.	1.	1.	0.	1.	0.	2.	1.	2.	3.
K/RB	717.	977.	1307.	102.	650.	766.	480.	672.	700.	1278.	1027.	1703.	778.	901.	585.	596.
K/SR	26.	13.	12.	7.	14.	6.	20.	9.	13.	47.	40.	91.	16.	24.	15.	15.
BA/RB	6.5	20.4	38.1	4.5	11.2	32.3	11.4	9.5	11.0	10.5	15.9	21.4	14.7	26.0	15.0	20.0
RB/SR	0.036	0.013	0.010	0.071	0.021	0.008	0.042	0.014	0.018	0.037	0.039	0.053	0.021	0.026	0.025	0.026
SR87/86	.0	.7039	.7036	.0	.0	.0	.0	.0	.7032	.0	.0	.7047	.0	.0	.7032	.7038
TH	ND	.63	.80	.56	.88	ND	1.2	.10	.30	ND	ND	ND	.31	ND	.81	ND
U	ND	.24	.17	.16	.31	ND	.61	.10	.17	.37	ND	ND	.26	ND	.33	ND
TH/U	0.0	2.6	4.7	3.5	2.9	0.0	2.0	1.0	1.8	0.0	0.0	0.0	1.2	0.0	2.4	0.0
K/U	0.	33566.	38394.	21030.	26011.	0.	26029.	28224.	20997.	0.	0.	0.	19157.	0.	15596.	0.
HPU	0.0	0.5	0.5	0.3	0.6	0.0	1.1	0.2	0.3	0.0	0.0	0.0	0.4	0.0	0.5	0.0
LA	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	4.0
YB	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	1.7
LA/YB	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	2.4
Y	24.	23.	24.	35.	22.	23.	29.	23.	25.	33.	23.	22.	28.	28.	27.	19.
NI	94.	20.	12.	15.	10.	12.	73.	23.	10.	21.	5.	6.	4.	3.	8.	ND
CO	46.	40.	34.	43.	35.	36.	40.	32.	32.	50.	18.	32.	23.	28.	26.	16.
CR	371.	32.	7.	20.	8.	26.	142.	105.	51.	94.	8.	16.	6.	5.	11.	11.
SC	36.	39.	36.	42.	29.	39.	39.	35.	35.	19.	30.	39.	39.	38.	27.	13.
V	235.	304.	265.	338.	204.	147.	275.	270.	270.	345.	309.	272.	199.	254.	182.	94.
CU	34.	91.	83.	56.	129.	37.	150.	69.	99.	65.	68.	78.	22.	108.	31.	24.
NI/CO	2.0	0.5	0.3	0.3	0.3	0.3	1.8	0.7	0.3	0.4	0.3	0.2	0.2	0.1	0.3	0.0
V/NI	3.	15.	23.	23.	20.	12.	4.	12.	27.	17.	60.	43.	49.	82.	22.	0.
MG/NI	309.	1889.	2764.	3115.	3269.	2488.	561.	1300.	2756.	1417.	3957.	4735.	5036.	6695.	3016.	0.
SN	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.94
HF	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	1.1
ZR	93.	65.	72.	64.	61.	21.	88.	45.	54.	85.	29.	105.	59.	36.	94.	56.
B	13.	27.	33.	0.	8.	10.	0.	0.	0.	0.	8.	15.	0.	10.	0.	0.
ZR/HF	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	49.
TI/ZR	78.	104.	98.	147.	105.	231.	74.	131.	119.	88.	163.	43.	105.	122.	50.	63.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

TABLE4.1. FIRST PERIOD VITI LEVU VOLCANICS.

DATA	386	52	366	906	48	866	812	913	55	60	914	915	857	903	901	856
SIO2	56.99	56.97	57.11	57.18	57.46	57.78	58.33	58.55	58.85	59.22	59.29	60.66	63.02	63.63	68.63	70.71
TIO2	0.64	1.81	0.75	1.28	0.64	1.14	0.85	1.05	1.28	1.08	1.09	1.11	0.93	0.96	1.04	0.38
AL203	19.61	16.90	17.97	16.53	19.57	15.41	15.07	15.01	16.00	15.13	15.19	15.93	16.42	16.23	15.53	15.42
FE203	7.65	4.86	7.42	10.37	2.99	10.94	9.70	11.82	3.57	8.47	8.84	8.11	6.04	6.14	3.59	3.87
FEO	0.0	4.24	0.0	0.0	4.23	0.0	0.0	0.0	5.13	2.54	0.0	0.0	0.0	0.0	0.0	0.0
MNO	0.15	0.25	0.16	0.20	0.15	0.26	0.19	0.17	0.22	0.15	0.24	0.14	0.19	0.23	0.09	0.09
MGO	2.50	3.46	4.03	3.54	2.30	3.41	3.73	3.35	3.46	3.50	2.39	2.85	1.95	1.81	0.28	0.80
CAO	8.54	3.50	8.56	4.07	8.74	4.26	8.02	6.24	6.29	6.17	9.21	7.38	4.74	4.09	3.90	2.41
NA20	2.97	7.00	3.38	6.05	3.02	6.60	3.42	3.26	4.30	3.32	2.72	2.87	4.44	4.82	5.30	5.33
K20	0.84	0.41	0.38	0.59	0.79	0.06	0.57	0.41	0.44	0.30	0.35	0.46	1.93	1.73	1.21	0.87
P205	0.10	0.61	0.25	0.20	0.10	0.16	0.12	0.14	0.45	0.13	0.69	0.48	0.33	0.37	0.43	0.11
(LOI)	2.33	3.54	1.86	2.79	2.21	2.09	5.50	3.86	1.99	4.24	2.91	3.90	4.35	2.86	0.42	3.25
(TOTAL)	99.53	99.38	99.57	99.12	99.84	99.57	99.49	98.86	99.59	99.44	100.02	98.86	99.03	99.55	98.99	99.68
NA20/K20	3.5	17.2	8.9	10.2	3.8	107.2	5.9	7.9	9.8	10.9	7.8	6.2	2.3	2.8	4.4	6.1
MG/MG+FE2	0.44	0.47	0.57	0.45	0.42	0.43	0.48	0.41	0.48	0.43	0.40	0.46	0.44	0.42	0.16	0.33
%K	0.70	0.34	0.31	0.49	0.65	0.05	0.48	0.34	0.37	0.25	0.29	0.38	1.60	1.43	1.00	0.72
RB	7.2	3.1	3.8	6.6	7.2	.62	5.7	7.4	4.1	2.2	7.2	3.5	50.	32.	16.	6.3
SR	174.	136.	604.	173.	172.	83.	240.	139.	224.	134.	171.	168.	393.	265.	277.	203.
BA	108.	52.	409.	269.	112.	16.	117.	62.	51.	74.	65.	49.	242.	350.	305.	214.
PB	2.	1.	4.	2.	4.	1.	1.	2.	0.	2.	1.	1.	4.	4.	2.	2.
K/RB	972.	1079.	830.	739.	913.	830.	830.	463.	892.	1146.	403.	1107.	323.	447.	617.	1143.
K/SR	40.	25.	5.	28.	38.	6.	20.	25.	16.	19.	17.	23.	41.	54.	36.	36.
BA/RB	15.0	16.7	108.1	40.5	15.6	26.7	20.4	8.4	12.5	33.3	9.0	14.2	4.9	10.9	18.8	33.8
RB/SR	0.041	0.023	0.006	0.038	0.042	0.007	0.024	0.053	0.018	0.016	0.042	0.021	0.126	0.121	0.059	0.031
SR87/86	.0	.7032	.0	.0	.7039	.0	.0	.0	.7035	.7037	.0	.7041	.0	.7034	.0	.0
TH	ND	.21	1.8	ND	ND	.21	ND	.11	ND	.21	ND	ND	1.7	ND	1.3	.81
U	ND	.21	.68	ND	ND	.21	.21	.11	ND	.11	ND	.36	.64	ND	1.3	.48
TH/U	0.0	1.0	2.7	0.0	0.0	1.0	0.0	1.0	0.0	2.0	0.0	0.0	2.6	0.0	1.0	1.7
K/U	0.	16188.	4654.	0.	0.	2490.	0.	32375.	0.	24074.	0.	0.	24904.	0.	7599.	15159.
HPU	0.0	0.3	0.9	0.0	0.0	0.2	0.0	0.2	0.0	0.2	0.0	0.0	1.2	0.0	1.5	0.7
LA	0.0	6.2	0.0	0.0	2.4	0.0	0.0	0.0	0.0	2.5	0.0	0.0	0.0	0.0	0.0	0.0
YB	.0	1.8	.0	.0	2.7	.0	.0	.0	.0	2.6	.0	.0	.0	.0	.0	.0
LA/YB	0.0	3.5	0.0	0.0	0.9	0.0	0.0	0.0	0.0	1.0	0.0	0.0	0.0	0.0	0.0	0.0
Y	27.	43.	20.	48.	29.	37.	36.	38.	0.	36.	40.	55.	69.	62.	63.	47.
NI	ND	ND	9.	4.	ND	5.	7.	ND	ND	ND	5.	ND	ND	ND	ND	ND
CO	17.	8.	21.	17.	18.	22.	23.	32.	16.	28.	21.	19.	10.	8.	ND	ND
CR	3.	ND	5.	ND	ND	2.	15.	5.	2.	2.	10.	6.	4.	4.	4.	2.
SC	21.	24.	18.	31.	29.	32.	36.	43.	27.	37.	44.	46.	17.	18.	18.	6.
V	121.	77.	184.	60.	154.	280.	246.	247.	154.	200.	292.	240.	70.	39.	43.	5.
CU	36.	2.	147.	189.	32.	111.	106.	88.	24.	81.	37.	33.	6.	11.	9.	3.
NI/CO	0.0	0.0	0.4	0.3	0.0	0.2	0.3	0.0	0.0	0.0	0.3	0.0	0.0	0.0	0.0	0.0
V/NI	0.	0.	20.	14.	0.	55.	33.	0.	0.	0.	57.	0.	0.	0.	0.	0.
MG/NI	0.	0.	2641.	5142.	0.	4005.	3024.	0.	0.	0.	2799.	0.	0.	0.	0.	0.
SN	.0	1.0	.0	.0	.80	.0	.0	.0	.0	1.4	.0	.0	.0	.0	.0	.0
HF	.0	.86	.0	.0	1.6	.0	.0	.0	.0	1.6	.0	.0	.0	.0	.0	.0
ZR	53.	107.	93.	122.	56.	101.	62.	69.	0.	71.	44.	53.	219.	208.	177.	162.
B	0.	0.	8.	0.	0.	0.	21.	0.	0.	0.	0.	0.	0.	0.	0.	0.
ZR/HF	0.	126.	0.	0.	34.	0.	0.	0.	0.	45.	0.	0.	0.	0.	0.	0.
TI/ZR	71.	101.	48.	62.	68.	68.	83.	91.	0.	91.	148.	126.	25.	28.	35.	14.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

TABLE4.1. FIRST PERIOD VITI LEVU VOLCANICS.

DATA	387	50	51	854
SIO2	72.12	73.31	73.45	74.26
TIO2	0.43	0.51	0.51	0.36
AL203	14.59	13.13	13.07	13.92
FE203	2.91	1.59	1.89	2.33
FEO	0.0	2.09	1.81	0.0
MNO	0.11	0.12	0.12	0.11
MGO	1.19	0.78	0.68	0.48
CAO	1.21	2.65	2.70	0.58
NA20	7.08	4.78	4.54	6.49
K20	0.27	0.91	1.11	1.40
P205	0.09	0.12	0.11	0.07
(LOI)	0.63	1.79	1.83	1.17
(TOTAL)	99.04	99.78	99.67	99.85
NA20/K20	25.8	5.3	4.1	4.6
MG/MG+FE2	0.50	0.32	0.30	0.33
%K	0.23	0.75	0.92	1.16
RB	3.3	6.9	9.1	15.
SR	91.	118.	121.	120.
BA	55.	222.	285.	299.
PB	2.	1.	1.	2.
K/RB	700.	1086.	1017.	764.
K/SR	25.	64.	77.	97.
BA/RB	16.9	32.1	31.3	19.7
RB/SR	0.036	0.059	0.075	0.127
SR87/86	.7045	.7047	.7041	.0
TH	.0	.0	.0	2.5
U	.0	.20	.20	1.0
TH/U	0.0	0.0	0.0	2.5
K/U	0.	0.	0.	11456.
HPU	0.0	0.0	0.0	1.6
LA	0.0	2.8	0.0	0.0
YB	.0	6.1	.0	.0
LA/YB	0.0	0.4	0.0	0.0
Y	44.	44.	45.	68.
NI	ND	ND	0.	ND
CO	ND	ND	3.	ND
CR	3.	ND	2.	3.
SC	6.	16.	18.	7.
V	8.	9.	9.	7.
CU	9.	ND	2.	4.
NI/CO	0.0	0.0	0.0	0.0
V/NI	0.	0.	0.	0.
MG/NI	0.	0.	0.	0.
SN	.0	3.0	.0	.0
HF	.0	4.6	.0	.0
ZR	189.	72.	74.	224.
B	0.	0.	0.	0.
ZR/HF	0.	16.	0.	0.
TI/ZR	14.	42.	42.	10.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

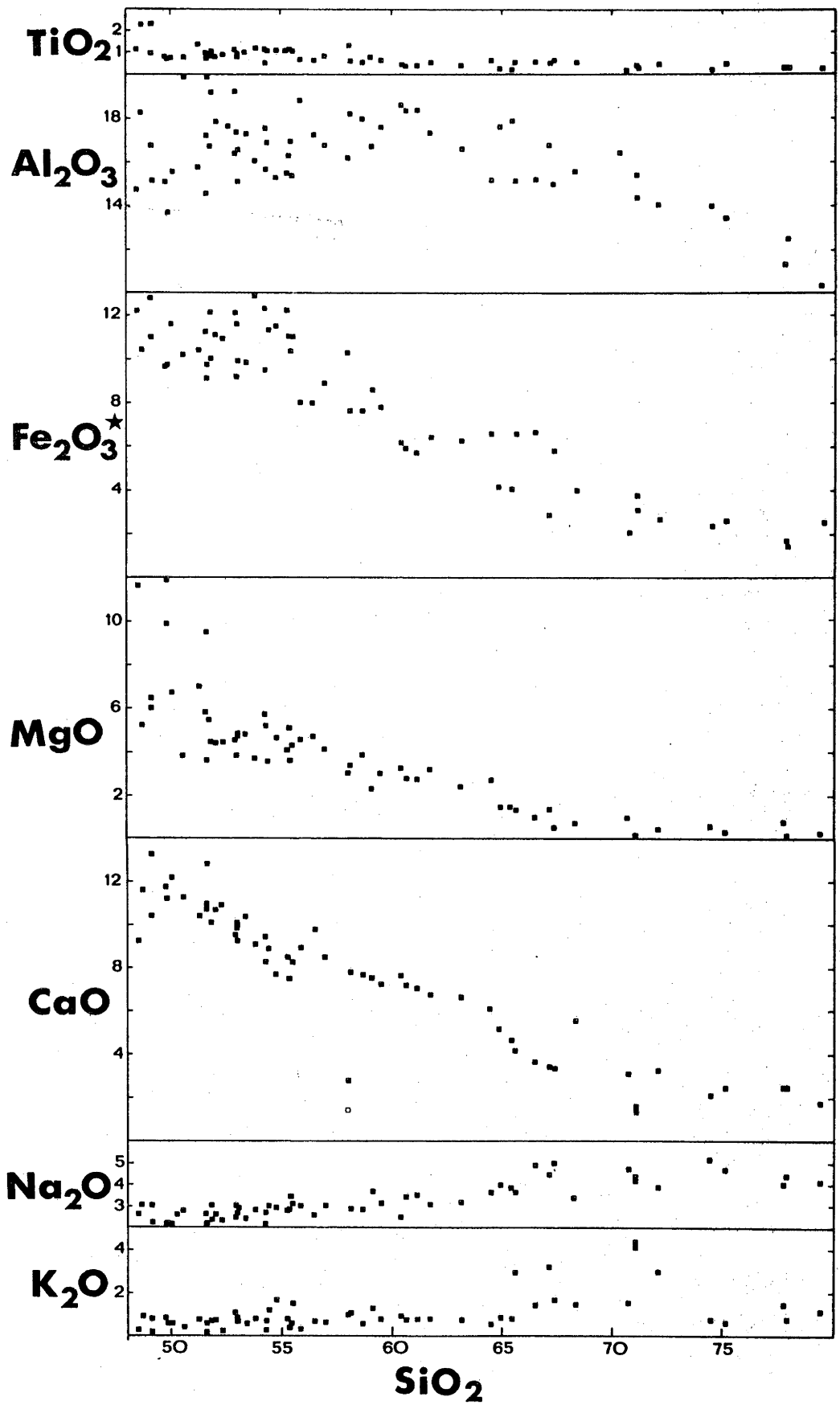

Figure 4.2 Major element oxides of first period, Viti Levu samples. Data from Table 4.1 and Rodda (1969). The open square indicates the Na_2O content of a highly albitized sample. Fe_2O_3* indicates total Fe as Fe_2O_3 as it also will in subsequent figures.

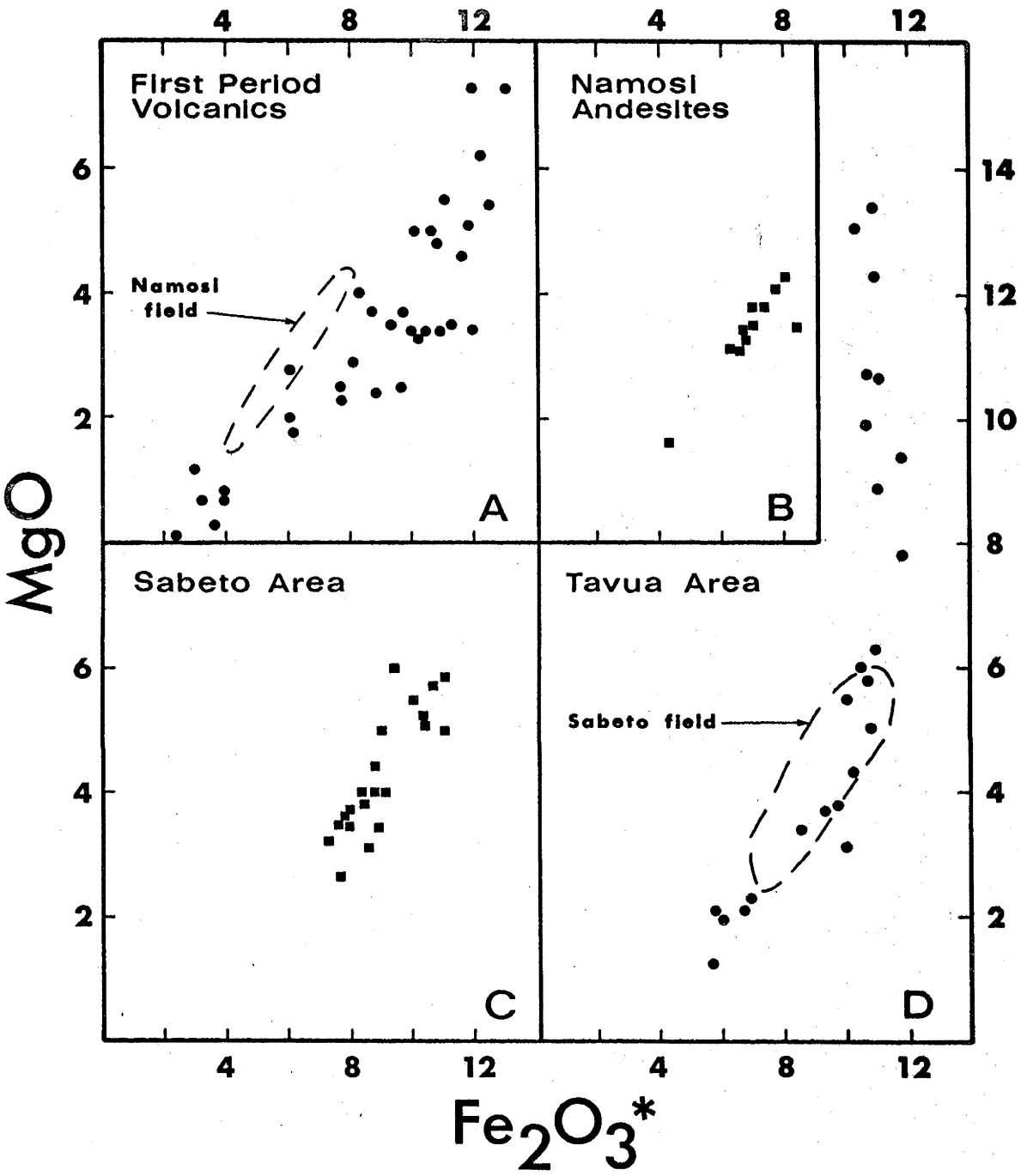

Figure 4.3 Fe-Mg variations in Viti Levu samples. Stratigraphically unambiguous samples from Tables 4.1, 4.3, 4.5, 4.6, Rodda (1969), and Ibbotson (1967). Note the difference between Namosi and first period volcanic rock fields, and the similarity between Sabeto and Tavua fields.

Sn, Zr, and Hf contents increase with SiO_2 whereas Ti decreases. Zr/Hf ratios are too erratic to judge whether deviations from their mean of 54 are significant.

Table 4.2 Rare earth analyses of first period, Viti Levu samples and a list of enrichment factors (e.f.) used

	52	54	48	60	50	e.f.
La	5.9	3.9	2.3	2.4	2.7	0.30
Ce	18.	10.	4.6	7.2	7.2	0.79
Pr	3.0	1.8	1.1	1.4	2.1	0.12
Nd	17.	9.1	7.0	8.4	14.	0.58
Sm	3.9	2.8	3.5	3.5	6.5	0.19
Eu	1.9	1.1	1.1	1.2	2.2	0.074
Gd	2.4	1.7	2.0	1.9	4.5	0.285
Tb	.49	.35	.51	.52	1.1	0.049
Dy	3.3	2.3	3.4	3.5	7.3	0.31
Ho	.76	.55	.90	.89	1.9	0.073
Er	1.5	1.3	2.1	1.8	4.9	0.21
Tm	.18	.17	.20	.23	.64	0.030
Yb	1.7	1.6	2.6	2.5	6.0	0.17

Table 4.2 contains REE data for five first period volcanics. Although poorer in quality than subsequent work, these data nevertheless define gross features. The pattern is generally flat relative to chondrites with maximum enrichment of Sm and Eu; La/Sm and La/Yb ratios are low as in ocean floor basalts. Heavy REE and Y contents increase with SiO_2. Samples 52 and 54, however, have higher light REE contents than do samples 48 and 60 although the former two are less fractionated from basaltic compositions. The Gd analyses are poor, making detection of Eu anomalies difficult. Only in aphyric sample 52, however, does Eu appear enriched; nowhere is it depleted.

Fifteen 87Sr/86Sr ratios were measured giving a mean of 0.7038. The range (0.7032 to 0.7047) well exceeds experimental error (0.0001-2) and indicates modest isotopic heterogeneity within the suite. There is a significant positive correlation (r = .64) between Rb/Sr and 87Sr/86Sr ratios (see Gill and Compston, Appendix 3, Figure 3a). Samples 50 and 51 are from the same site and nearly isochemical except for their Sr isotopic composition. If the analytically less reliable

isotope data for sample 51 is omitted, the correlation is enhanced ($r = .74$) and becomes significant at the 99.75% confidence level. Both remaining rhyolites have high ratios (0.7045 and 0.7047). The equally high ratio for sample 49 was reconfirmed and unaffected by HCl leaching. Similar leaching of sample 60 showed the 0.7049 ratio reported earlier for it (Gill, 1970) to be due to HCl-soluble Sr; the 87Sr/86Sr ratio of the insoluble residue is 0.7037.

D. Geochemical effects of submarine weathering and burial metamorphism

As mentioned above, these volcanics are not pristine, virginal rocks. They were largely submarine initially and many have experienced some degree of burial metamorphism. Both processes could have modified rock compositions. Indeed, most samples have high LOI (which underestimates H_2O+CO_2+other volatiles) and high Fe_2O_3 (>3%) in basalts and andesites.

Submarine weathering has been studied in many rocks dredged from the ocean floor and its effects found to be important in those rocks having >0.7% H_2O and Fe_2O_3/FeO ratios >0.3. The freshest submarine basalts have $H_2O^+/K_2O \sim 1$ and both <0.25%, and <1.5% Fe_2O_3 (Hart and Nalwalk, 1970; Hart, 1971a; Moore, 1965; Hekinian, 1971). Sea water alteration results in increased H_2O^+, K, Rb, Cs, B, Li, and U contents and Fe_2O_3/FeO and K/Rb ratios (Hart, 1969, 1971a; Cann, 1970; Hekinian, 1971; Thompson and Melson, 1970; Aumento, 1971). Increases can be by factors >10 for Cs, B, and U, and about two to eight for Rb. Sr isotope ratios may also increase (Hart and Nalwalk, 1970; Pushkar and Peterson, 1967) approaching that of sea water (0.709). In contrast, REE, Y, Zr, Nb, and Ti are not affected (Philpotts et al., 1969; Cann, 1970; Hekinian, 1971). Ba and Sr contents vary randomly but within a factor of two or less.

Some effects of burial metamorphism are similar to those of submarine weathering. However, metamorphism to no more than greenschist facies assemblages can be almost a closed-system phenomenon, although the scale for which this is true varies with chemical species (see Liou and Kennedy, 1971). In every published study, such metamorphism is accompanied by increased H_2O^+, CO_2, and Fe_2O_3/FeO (e.g. Cann, 1969; Herrmann and Wedepohl, 1970; Hekinian, 1971; Miyashiro et al., 1971; Levi, 1969; Aumento et al., 1971). Replacement of plagioclase by albite and clinopyroxene by chlorite

releases Ca which, in the Viti Levu instance, was reprecipitated as calcite, epidote, sphene, or Ca-zeolites. Local enrichments in Na and depletions in Ca and Si are commonly reported (Cann, 1969; Smith, 1968; Herrmann and Wedepohl, 1970; Levi, 1969; Aumento et al., 1971). Na_2O and H_2O^+ may vary conconcomitantly and Na_2O/K_2O ratios increase. Once again REE, Y, Zr, Nb, and Ti are unaffected (Cann, 1970; Herrmann and Wedepohl, 1970; Frey et al., 1968; Graham and Nicholls, 1969). B apparently does not increase (Thompson and Melson, 1970) and Sr falls only when Ca does (Cann, 1969). Ni, Co, Cr, Sc, and V seem well accomodated in chlorite and their concentrations vary little (Cann, 1969; Melson et al., 1968) although Cu can be quite mobile depending on the composition of the fluid phase. Alkali behavior is erratic: Cann (1969, 1970) and Aumento et al. (1971) found less K and Rb in greenschists than fresh ocean floor basalts whereas Matthews (1971) and Levi (1969) found marked K_2O enrichments.

Thus, elements such as Ti, Zr, Nb, REE, B, Sr, and the transition metals are unlikely to be affected on a 1-2 kg sample size scale. The alkalis, Ca, and Si can be more mobile in the accompanying aqueous phase--depleted locally and enriched elsewhere.

Some features of first period volcanic rock compositions can be attributed to these alteration processes. High Na_2O accompany low CaO contents in rocks with modal albite and/or very turbid plagioclase. Usually LOI and Fe_2O_3/FeO ratios are also high. The low concentration of most large cations, however, seems to be a primary feature although irregularities could result from either alteration process. K, P, Rb, Ba, Pb, and Sr are consistently low in both fresh and altered samples. Specific enrichments or depletions relative to other rocks of the suite may reflect local mobility of these cations. As mentioned above, HCl-soluble Sr in sample 60 had a higher 87/86 ratio than the insoluble Sr, demonstrating mobility of Sr without isotopic equilibration. Such Sr could have been dissolved from neighboring rocks with higher Rb/Sr or be from connate sea water, and precipitated with Ca in calcite or zeolite.

U and Th were rarely detected in metamorphosed rocks although they were measureable in fresher samples (60, 904, 920). U, particularly 234U, is often soluble in oxidizing aqueous solutions although Th is less so (Cherdyntsev et al., 1964; Rosholt et al., 1964, 1966) which might explain their very low concentrations in

greenschists. However, 234U/238U ratios average 1.1 for analyses of seven imprecisely-defined but probably first period volcanic rocks from Viti Levu (Cherdyntsev and Senina, 1970), precluding selective 234U solubility. The average U content and Th/U ratio in Cherdyntsev and Senina's six "basalts" and one "andesite" are 0.2 ppm and 2, respectively. Th and U retain significant correlation (r = .85) in my samples, also mitigating against U leaching. U and Th contents and Th/U ratios are therefore considered to have been inherently low but possibly affected by metamorphism.

Other element abundances and ratios should have been relatively unaffected by these alteration processes.

E. Summary

Most features summarized in section 4.2C are considered primary geochemical characteristics of first period volcanic rocks. Salient aspects are the wide Si range; low Ti and Zr; variable Al; relatively high Fe and Fe/Mg; low K, Rb, Ba, Pb, Sr, P, U, and Th; relatively flat REE patterns with low La/Sm and La/Yb; low Ni and Cr; and 87Sr/86Sr ratios which average 0.7038 but which are positively correlated with Rb/Sr. Modest Fe-enrichment is especially noticeable at 57-59% SiO_2 and $3\frac{1}{2}$% MgO, and is accompanied by increased TiO_2. K, Rb, and Ba are somewhat enriched in a few dacites but little more abundant in andesites or rhyolites than basalts. Heavy REE and Y increase with Si contents.

Some exceptions to this summary merit comment. Samples 61, 62, and 367 are from Nasinu quarries near Suva which expose rocks thought to pre-date the Middle Miocene unconformity (Ibbotson, 1960). They have more K, Sr, and B and less Ca and Mg than other first period basalts and apparently represent a genetically somewhat different tradition. Considerable scatter is evident in Figure 4.3a and in the Fe_2O_3* portion of Figure 4.2, implying that the Fe-enrichment process was inconsistent. Low-Fe sample 366 is a stream cobble from near outcrops of Namosi Andesite rocks to which it bears strong geochemical resemblance except in K, Rb, U, and Th. Similarly, low-Fe samples N315 and N396 in Rodda (1969) are dikes intruding Wainimala Group rocks in southwestern Viti Levu and could also be related to later volcanism. Sample 907 is from an outcrop mapped as belonging to the Wainimala Group but the outcrop, thin section, and analysis seem more like Namosi than Wainimala except that the rock is an ol-normative

basalt. I interpret the variable alkali contents in analyzed dacites as reflecting different geneses although alteration may be responsible for the enrichments.

4.3 Second period volcanism: Late Miocene

A. Introduction

Volcanic rocks are associated with Tertiary g faunas in the Mendrausuthu Andesitic and Nandi Sedimentary Groups (Ibbotson, 1960; Bartholomew, 1959a; Rickard, 1963) which unconformably overlie first period rocks. The former group also includes planktonic foraminifera of zones N.17-19 (Eames et al., 1958; Blow, 1969) and detritus from 7-13 m.y. old plutonics (Rodda, 1967). The age of these second period volcanics is therefore thought to be Late Miocene to Early Pliocene. No rocks of the Nandi Sedimentary Group were analyzed.

The geologic setting and petrography of the Namosi Andesite member of the Mendrausuthu Andesitic Group have been discussed by Ibbotson (1960), Band (1968), and Rodda (1967). (The most detailed study is a bulletin by P. Rodda, written but as yet unpublished and to which I have not had access.)

B. Petrography

Augite and plagioclase are the dominant phenocrysts, frequently accompanied by a few per cent magnetite and joined, higher in the stratigraphic sequence (Rodda, 1967), by hornblende. Phenocrysts typically comprise about 50% of the mode. Analyses of mafic phenocrysts from a Mau Quarry sample (Jakes, 1970) are given in Table 10.2. Plagioclase phenocrysts usually have basic labradorite to bytownite cores, complex oscillatory zoning, and ubiquitous inclusions of glass, clinopyroxene, and magnetite. Augites are fresh, zoned, and occasionally have plagioclase inclusions. Hornblendes usually have strong reaction rims. The groundmass is normally fresh with quench crystallites of plagioclase, pyroxene, and an opaque. Glass usually accompanies pyroxene but rarely amphibole phenocrysts.

C. Geochemistry

Six new analyses are presented in Tables 4.3 and 4.4 and Figures 4.3b, 4.4 and 4.5b together with five others from Jorgensen (in Band, 1968), Taylor et al., (1969), Gill (1970), and Jakes (1970). The REE, Th, and U data for samples X88 and X96 represent new analyses of Taylor et al.'s rock powders. In section 4.5A I will discuss my reasons for including sample 870 from the Wainikoro plutonic stock

TABLE 4.3. SECOND PERIOD VITI LEVU VOLCANICS.

DATA	869	876	868	870	873	X96	PJ6	PJ18	66	X88	874
SIO2	55.34	55.43	56.54	56.60	57.97	58.60	59.62	59.62	60.09	60.17	65.45
TIO2	0.89	0.82	0.77	0.73	0.70	0.66	0.67	0.70	0.69	0.71	0.52
AL203	17.81	18.05	17.50	18.00	17.61	18.07	17.12	17.24	16.93	17.26	16.59
FE203	8.44	8.03	7.73	7.43	7.04	3.27	3.83	3.94	3.70	2.98	4.29
FEO	0.0	0.0	0.0	0.0	0.0	3.37	2.60	2.59	2.64	2.93	0.0
MNO	0.18	0.13	0.15	0.12	0.13	0.0	0.13	0.15	0.13	0.26	0.09
MGO	3.52	4.26	4.14	3.84	3.84	3.52	3.44	3.28	3.12	3.18	1.59
CAO	8.85	8.36	8.17	7.49	7.51	7.55	7.03	7.16	7.20	7.32	5.17
NA20	3.70	3.42	3.73	4.26	3.73	3.47	3.97	3.84	4.02	3.79	4.60
K20	1.02	1.28	1.03	1.28	1.24	1.28	1.38	1.28	1.28	1.26	1.45
P205	0.24	0.23	0.26	0.24	0.23	0.21	0.20	0.20	0.20	0.14	0.25
(LOI)	1.60	2.42	1.17	1.18	1.53	1.60	0.85	0.64	0.55	0.67	0.42
(TOTAL)	99.21	99.32	99.02	99.46	99.01	99.56	99.56	99.68	99.39	99.72	99.24
NA20/K20	3.6	2.7	3.6	3.3	3.0	2.7	2.9	3.0	3.1	3.0	3.2
MG/MG+FE2	0.50	0.56	0.57	0.56	0.57	0.55	0.55	0.54	0.53	0.55	0.47
%K	0.85	1.06	0.86	1.06	1.03	1.06	1.14	1.06	1.07	1.05	1.20
RB	14.	18.	17.	16.	36.	22.	23.	19.	16.	22.	28.
SR	536.	553.	622.	588.	550.	551.	517.	506.	487.	449.	660.
BA	318.	426.	377.	392.	437.	276.	562.	537.	467.	333.	417.
PB	3.	4.	5.	2.	4.	0.	0.	0.	4.	2.	4.
K/RB	593.	606.	493.	654.	287.	472.	491.	555.	659.	472.	424.
K/SR	16.	19.	14.	18.	19.	19.	22.	21.	22.	23.	18.
BA/RB	22.1	24.3	21.7	24.1	12.2	12.3	24.1	28.0	28.9	15.0	14.7
RB/SR	0.027	0.032	0.028	0.028	0.065	0.041	0.045	0.038	0.033	0.049	0.043
SR87/86	.0	.7038	.0	.0	.7036	.0	.0	.0	.7039	.0	.7035
TH	1.4	1.5	1.9	1.7	1.6	1.4	1.8	1.3	2.0	1.9	1.5
U	.52	.53	.60	.58	.69	.59	.76	.0	.72	.76	.82
TH/U	2.7	2.9	3.2	3.0	2.4	2.4	2.4	0.0	2.8	2.6	1.9
K/U	16277.	20184.	14211.	18350.	14992.	17891.	15053.	0.	14849.	13835.	14655.
HPU	0.9	1.0	1.1	1.1	1.1	1.0	1.2	0.0	1.2	1.2	1.2
LA	0.0	8.9	0.0	0.0	0.0	11.3	0.0	0.0	0.0	9.2	10.6
YB	.0	1.4	.0	.0	.0	1.4	.0	.0	.0	1.4	1.4
LA/YB	0.0	6.1	0.0	0.0	0.0	7.9	0.0	0.0	0.0	6.5	7.5
Y	27.	22.	25.	21.	25.	18.	18.	18.	17.	20.	19.
NI	5.	7.	13.	7.	8.	0.	9.	9.	11.	8.	7.
CO	26.	23.	25.	18.	19.	0.	0.	0.	16.	16.	10.
CR	10.	8.	43.	8.	10.	0.	46.	42.	31.	36.	6.
SC	25.	26.	21.	18.	19.	0.	0.	0.	14.	18.	8.
V	256.	283.	220.	197.	191.	0.	196.	159.	172.	162.	85.
CU	144.	44.	101.	88.	181.	0.	42.	29.	44.	30.	8.
NI/CO	0.2	0.3	0.5	0.4	0.4	0.0	0.0	0.0	0.7	0.5	0.7
V/NI	50.	39.	17.	28.	23.	0.	21.	17.	15.	20.	12.
MG/NI	4150.	3559.	1879.	3248.	2820.	0.	2279.	2178.	1689.	2375.	1353.
SN	.0	.21	.0	.0	.0	.37	.0	.0	.0	.29	.26
HF	.0	1.2	.0	.0	.0	1.3	.0	.0	.0	1.2	1.6
ZR	90.	87.	100.	82.	101.	97.	0.	0.	112.	116.	119.
B	7.	11.	18.	5.	29.	0.	0.	0.	0.	0.	0.
ZR/HF	0.	70.	0.	0.	0.	73.	0.	0.	0.	96.	74.
TI/ZR	59.	56.	46.	53.	42.	41.	0.	0.	37.	36.	26.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

X88 AND X96 FROM TAYLOR ET AL.(1969); PJ6 AND PJ18 FROM JAKES(1970)

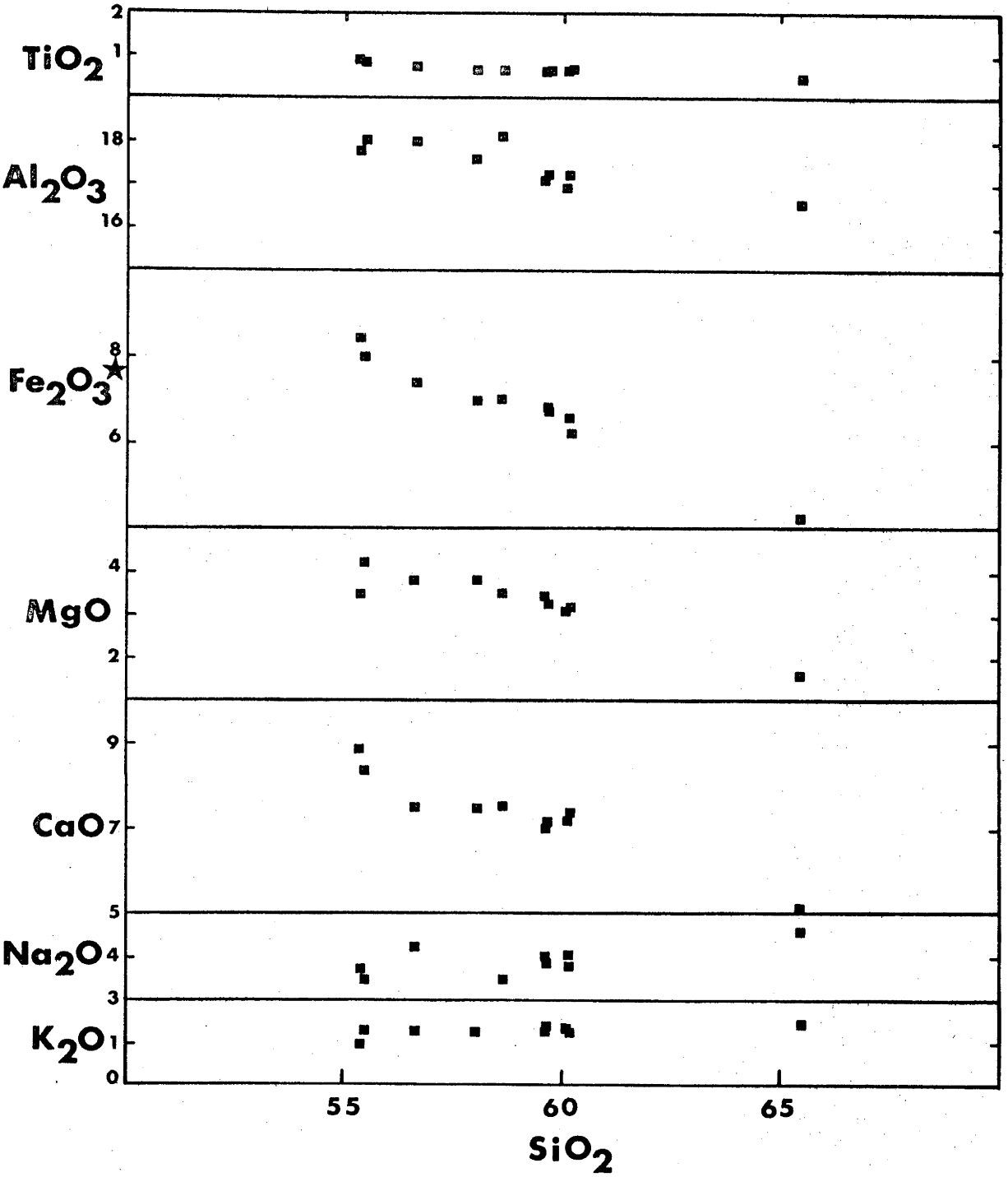

Figure 4.4 Major element oxides of Namosi Andesites, Viti Levu. Data from Table 4.3.

here rather than with the Tholo Plutonics.

The eleven analyses define a suite of intermediate silica content comprising basaltic-andesites, andesites, and one dacite. Na_2O/K_2O ratios decrease slightly from 3.6 to about 3 in more felsic samples. $Mg/Mg+Fe^{+2}$ ratios are similar in basaltic-andesites and andesites and Figure 4.3b illustrates the classically calc-alkaline feature of minimal iron enrichment within the suite. Trace element contents closely approximate those thought average for calc-alkaline andesites by Taylor (1969). Measured $^{87}Sr/^{86}Sr$ ratios have a narrow range (0.0004) and average 0.7037. Such characteristics justify my earlier description of these rocks as "typically calc-alkaline" (Gill, 1970) and were the reasons why T. Green chose one as his starting material for high pressure studies relevant to calc-alkaline magma genesis (T. Green, 1972). K/Rb ratios average 520; Th/U, 2.5; Ba/Rb, 20; and Zr/Hf, 78. Sn contents are low as are Nb (1.5, 0.9, and 3.3 ppm in samples 876, 66, and 874 respectively: XRF analyses by M. Kaye, 1971).

Table 4.4 Rare earth analyses of Namosi Andesites, Viti Levu

	876	X96	X88	874
La	8.6	11.1	9.1	10.5
Ce	17.8	22.2	20.0	22.2
Pr	2.4	2.9	2.7	3.5
Nd	11.0	12.2	11.3	14.4
Sm	2.6	2.4	2.4	3.3
Eu	.76	.68	.77	1.0
Gd	2.6	2.5	-	3.4
Tb	.39	.42	.44	.50
Dy	2.2	1.9	2.7	3.0
Ho	.50	.50	.48	.53
Er	1.4	1.3	1.2	1.4
Yb	1.4	1.4	1.4	1.4

The differences in composition between volcanic rocks of the first and second periods are considerable and important. In second period samples the SiO_2 range is contracted, Fe-enrichment is less with correspondingly higher $Mg/Mg+Fe^{+2}$ ratios, Na_2O/K_2O ratios are lower, and large cation minor and trace elements (K, P, Rb, Sr, Ba, Pb, Th,

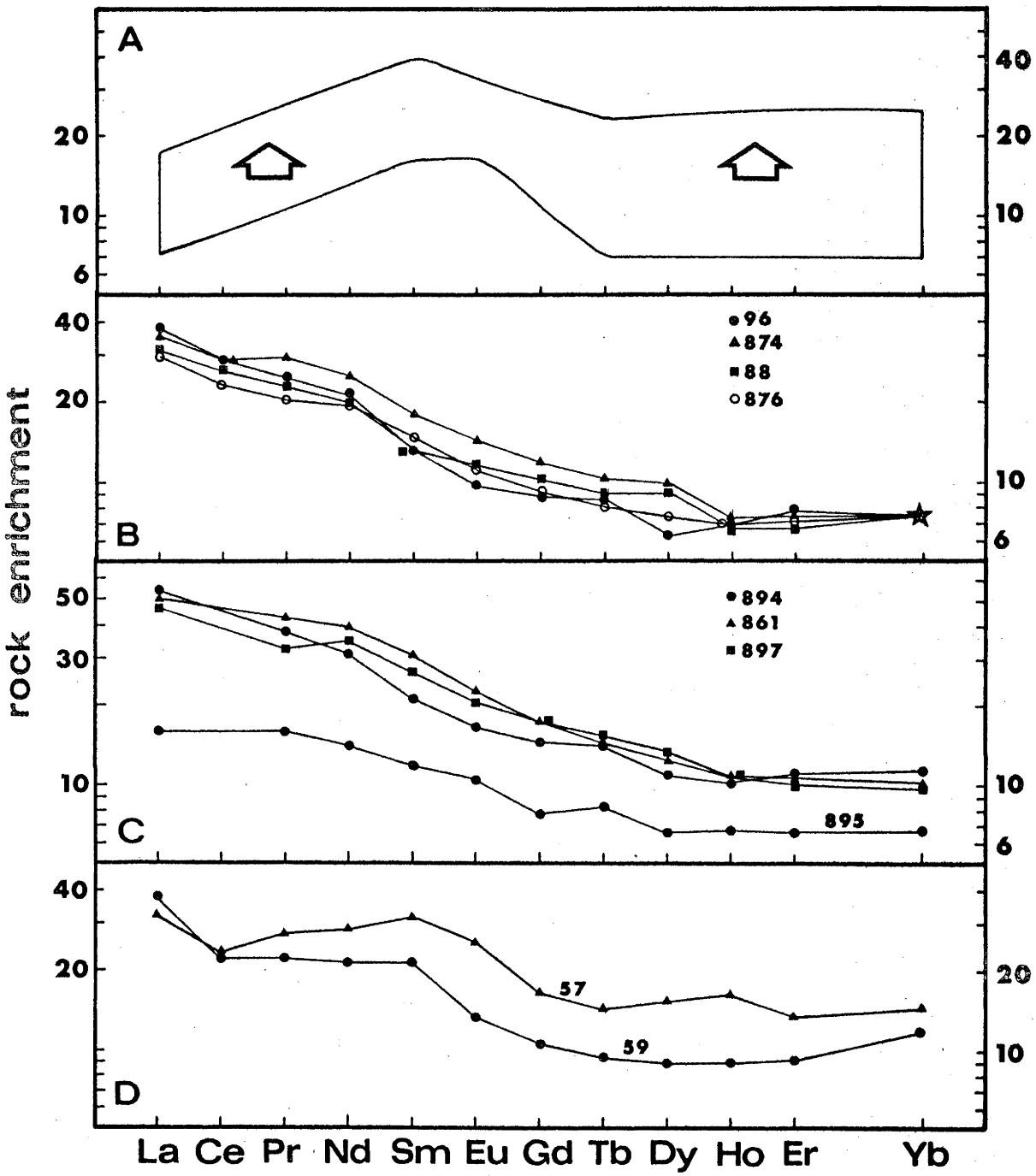

Figure 4.5 Rare earth patterns of Viti Levu samples. Data from Gill (1970) and Tables 4.4 and 4.9. The outline in area A indicates the range of REE patterns; arrows indicate the direction of change with increasing silica contents. The star in area B indicates the composition of all four samples.

and U) are two to four times more abundant. Th/U ratios are consistently higher than in most analyses of first period rocks; K/Rb and K/U ratios are lower. Four Namosi Andesite REE patterns define a narrow field inclined rather than parallel to those of chondrites (Figure 4.5b) with La/Yb ratios of six to eight. Heavy REEs are consistently less than ten times more abundant than in chondrites. Ni and Cr remain detectable at lower MgO contents in second period rocks; Mg/Ni ratios are lower and V/Ni ratios higher.

Thus the stratigraphic unconformity separating rocks of these two periods also represents a marked geochemical discontinuity.

4.4 Third period volcanism: Pliocene and younger

A. Introduction

Predictably the youngest rocks are the best preserved and most studied, and their complexity is most obvious. Several eruptive centers yielded these largely basic volcanic rocks. These are, in the order I shall discuss them, the Sabeto, Tavua, Rakiraki, Vatia, Nakobalevu, and Beqa centers which produced those rocks comprising the Koroimavua Andesitic and Mba Basaltic Groups. Dickinson et al. (1968) estimate 10^4 km^3 of "late Cainozoic shoshonite in Fiji", and others also suggest a considerable volume of these rocks. Rodda (1967, p. 1265) reports almost one km of exposure; Cohen (1962) estimates Mba rocks in the vicinity of Vatukoula to be 2.5 km thick; Robertson (1967) found Viti Levu's lowest Bouguer anomalies in this outcrop area; and Malahoff (preprint) interpreted the aero-magnetic data available to him as indicating basaltic source bodies at 1.5 to 5 km depths.

Sediments are associated with several centers and include a fauna that is Tertiary h or younger in age (Rodda, 1967; Rickard, 1963; Hirst, 1965). In places these sediments overlie rocks of the first and/or second periods unconformably (Rodda, 1967; Rickard, 1963; Hirst, 1965; Band, 1968; Dickinson, 1968b). McDougall (1963) found biotites from four Sabeto center samples to yield consistent ages between 4.8 and 5.3 m.y. and Rodda et al. (1967) report six similar results (4.3 to 5.0 m.y.) from the Tavua center and related rocks. They also report two older results from that crater (7.5 ±1 and 10 ±0.5 m.y.) and another from the Vatia center (9.2 ±2 m.y.), the significance of which is unclear without further documentation. Volcanism apparently began and, in the western centers at least, ended during the Pliocene.

Most rocks of this period are basaltic or, to be more precise, shoshonitic (see section 1.3 for an explanation of nomenclature).

B. Sabeto centers

I use this term to denote the volcanic rocks of both Koroimavua and Mba Groups and the intrusive stocks of northwestern Viti Levu, principally in the Sabeto Range overlooking Nadi International Airport. Rickard (1963), Dickinson (1968b and written comm., 1969), and others distinguish lithologically between Koroimavua and Mba Groups in this area although with difficulty (e.g. Rickard, 1963, pp. 20-25), using color index, density, and presence or absence of hornblende as diagnostic criteria. Chemically there is no distinction and I have therefore made none.

Zoned labradorite and augite are ubiquitous phenocrysts. Augite occurs without or exceeds plagioclase in some samples but is usually less abundant. Most modal and normative feldspar is labradorite. Olivine is apparently absent from Mba basalts in this area. Biotite occurs as phenocrysts (<5%) in some rocks with high K/Na and Al, and also in some matrices. Several per cent of oxides may also occur as phenocrysts. There were neither K-feldspar nor leucite phenocrysts in my volcanic samples although number 881, from an associated intrusive stock, is about 3/4 K-feldspar. (Leucite has been reported in Koroimavua rocks: Dickinson *et al.*, 1968; Rickard, 1963). Phenocryst minerals, the plagioclase sometimes mantled by K-feldspar, together with devitrifying glass and alteration products comprise the groundmass of most samples (Appendix 2 and Crook, 1963).

Previously Iyengar (in Rickard, 1963) reported one analysis of a rock from this area; Dickinson *et al.* (1968) three; Gill (1970) five; and Rodda (1969) four of associated intrusive stocks. Peterman (written comm., 1969) measured Sr, Rb, and 87Sr/86Sr ratios for the three samples of Dickinson *et al.* These data plus nine new analyses are summarized in Table 4.5 and Figure 4.6.

Despite the appellation "Koroimavua Andesitic Group", no specimen yet analyzed is an andesite in composition: all are much more mafic and most rocks of both Groups are absarokites of the shoshonitic rock suite. The SiO_2 range is narrow and somewhat high relative to most basaltic suites (49.5 to 53.2%); MgO contents and $Mg/Mg+Fe^{+2}$ ratios (<6% and <.62, respectively) are low by such comparisons. Al_2O_3 and TiO_2 contents are similar to those of earlier Viti Levu and most

TABLE 4.5. THIRD PERIOD VITI LEVU VOLCANICS: SABETO AREA.

DATA	881	861	71	865	NA-3	75	LD39	884	BB-2	65	882	64	863	63	864	887	886
SIO2	50.38	48.79	50.08	50.27	50.35	50.70	50.72	51.14	51.51	51.69	51.76	51.78	51.80	52.03	52.20	52.74	53.25
TIO2	0.72	0.73	1.01	0.69	0.67	0.75	0.70	0.71	0.72	0.63	0.59	0.63	0.65	0.56	0.67	0.75	0.70
AL203	18.60	16.18	16.03	15.19	18.47	17.68	16.98	20.28	18.47	20.24	18.83	20.27	17.11	17.56	21.20	18.42	18.38
FE203	8.36	10.36	6.00	10.30	5.46	5.26	6.27	8.27	5.69	3.59	7.89	3.68	8.75	4.79	7.35	8.79	8.67
FEO	0.0	0.0	4.56	0.0	2.73	3.81	3.32	0.0	3.02	3.86	0.0	3.72	0.0	2.49	0.0	0.0	0.0
MNO	0.23	0.20	0.20	0.20	0.22	0.14	0.18	0.15	0.16	0.14	0.18	0.15	0.17	0.18	0.12	0.21	0.20
MGO	3.91	5.14	5.85	5.18	4.40	6.02	5.52	3.98	5.07	3.67	3.49	3.77	4.38	3.46	3.15	3.40	4.01
CAO	9.22	10.85	10.21	11.71	9.62	10.73	9.52	8.78	9.62	9.11	9.19	9.24	9.22	9.25	7.73	9.97	8.70
NA20	3.91	3.53	2.93	2.86	2.74	2.50	2.27	2.97	2.45	3.08	3.75	2.96	3.70	5.11	3.07	3.36	3.18
K20	4.17	3.57	2.75	2.81	4.84	2.16	3.95	3.27	2.91	3.52	3.88	3.34	3.76	4.12	3.91	1.93	2.54
P205	0.52	0.67	0.39	0.79	0.50	0.26	0.56	0.45	0.39	0.46	0.45	0.45	0.45	0.45	0.60	0.44	0.36
(LOI)	3.32	3.42	1.41	5.95	2.51	2.03	1.54	1.62	1.72	1.40	2.98	1.80	4.86	3.33	3.40	2.38	1.75
(TOTAL)	99.77	99.30	100.11	99.05	99.45	100.00	100.24	98.91	99.82	99.66	99.31	99.91	99.76	100.39	100.21	99.24	99.45
NA20/K20	0.9	1.0	1.1	1.0	0.6	1.2	0.6	0.9	0.8	0.9	1.0	0.9	1.0	1.2	0.8	1.7	1.3
MG/MG+FE2	0.53	0.55	0.56	0.55	0.56	0.61	0.57	0.54	0.58	0.53	0.52	0.54	0.55	0.53	0.51	0.48	0.53
%K	3.46	2.96	2.28	2.34	4.02	1.80	3.28	2.71	2.41	2.92	3.22	2.78	3.12	3.42	3.25	1.60	2.11
RB	85.	67.	30.	55.	52.	38.	59.	71.	88.	65.	39.	57.	70.	73.	93.	39.	57.
SR	1635.	1491.	1298.	1211.	2383.	658.	2168.	1100.	1386.	1257.	1667.	1252.	1707.	1599.	949.	930.	876.
BA	596.	584.	462.	956.	0.	510.	0.	596.	0.	699.	675.	690.	632.	698.	661.	330.	368.
PB	10.	8.	7.	9.	0.	5.	0.	6.	0.	8.	9.	8.	6.	11.	6.	6.	2.
K/RB	407.	444.	750.	426.	779.	476.	558.	383.	275.	449.	817.	486.	449.	468.	350.	409.	368.
K/SR	21.	20.	18.	19.	17.	27.	15.	25.	17.	23.	19.	22.	18.	21.	34.	17.	24.
BA/RB	7.0	8.7	15.2	17.5	0.0	13.5	0.0	8.4	0.0	10.7	17.1	12.1	9.1	9.5	7.1	8.4	6.4
RB/SR	0.052	0.045	0.023	0.045	0.022	0.057	0.027	0.064	0.063	0.052	0.024	0.046	0.041	0.046	0.098	0.042	0.065
SR87/86	.0	.0	.7043	.0	.7043	.7042	.7040	.0	.7038	.7040	.0	.7037	.0	.7043	.0	.0	.0
TH	2.1	2.1	1.0	2.7	.0	.53	.0	.75	.0	.87	2.2	1.3	2.2	2.3	.99	1.1	1.2
U	1.3	1.4	.50	1.8	.0	.18	.0	.29	.0	.55	1.2	.57	.89	1.2	.62	.52	.63
TH/U	1.5	1.5	2.0	1.5	0.0	2.9	0.0	2.6	0.0	1.6	1.8	2.3	2.5	1.8	1.6	2.2	1.9
K/U	25670.	21839.	45911.	12794.	0.	97771.	0.	94279.	0.	53190.	25872.	48622.	35281.	27671.	52437.	31047.	33205.
HPU	2.3	2.2	1.2	2.5	0.0	0.7	0.0	1.1	0.0	1.4	2.2	1.4	1.9	2.3	1.5	1.0	1.3
LA	0.0	15.6	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	8.7	0.0	0.0	0.0	0.0	0.0
YB	.0	1.9	.0	.0	.0	.0	.0	.0	.0	.0	.0	1.5	.0	.0	.0	.0	.0
LA/YB	0.0	8.3	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	5.7	0.0	0.0	0.0	0.0	0.0
Y	25.	27.	23.	29.	0.	16.	0.	21.	0.	13.	21.	12.	21.	13.	20.	28.	25.
NI	5.	17.	17.	25.	0.	26.	0.	14.	0.	13.	6.	13.	13.	6.	10.	5.	9.
CO	27.	32.	32.	39.	0.	32.	0.	24.	0.	23.	22.	21.	25.	21.	17.	25.	27.
CR	2.	18.	9.	35.	0.	73.	0.	32.	0.	79.	4.	41.	14.	3.	17.	6.	6.
SC	17.	30.	29.	33.	0.	35.	0.	17.	0.	16.	16.	14.	23.	15.	12.	19.	21.
V	294.	355.	355.	349.	0.	337.	0.	259.	0.	265.	254.	224.	274.	340.	171.	238.	240.
CU	240.	218.	101.	143.	0.	81.	0.	225.	0.	265.	230.	204.	203.	185.	264.	317.	49.
NI/CO	0.2	0.5	0.5	0.6	0.0	0.8	0.0	0.6	0.0	0.6	0.3	0.6	0.5	0.3	0.6	0.2	0.3
V/NI	57.	21.	21.	14.	0.	13.	0.	18.	0.	20.	41.	17.	22.	55.	17.	46.	26.
MG/NI	4548.	1859.	2047.	1264.	0.	1423.	0.	1667.	0.	1675.	3378.	1717.	2091.	3378.	1840.	3969.	2627.
SN	.0	.45	.0	.0	.0	.0	.0	.0	.0	.0	.0	1.6	.0	.0	.0	.0	.0
HF	.0	1.1	.0	.0	.0	.0	.0	.0	.0	.0	.0	1.1	.0	.0	.0	.0	.0
ZR	55.	56.	65.	107.	0.	27.	0.	43.	0.	47.	43.	47.	45.	0.	58.	61.	68.
B	11.	31.	0.	13.	0.	0.	0.	0.	0.	18.	22.	-33.	16.	18.	18.	3.	6.
ZR/HF	0.	49.	0.	0.	0.	0.	0.	0.	0.	0.	0.	42.	0.	0.	0.	0.	0.
TI/ZR	78.	78.	94.	38.	0.	168.	0.	98.	0.	81.	83.	81.	86.	0.	70.	74.	62.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

NA-3, LD39, AND BB-2 FROM DICKINSON ET AL.(1968) AND PETERMAN(WRITTEN COMM., 1969)

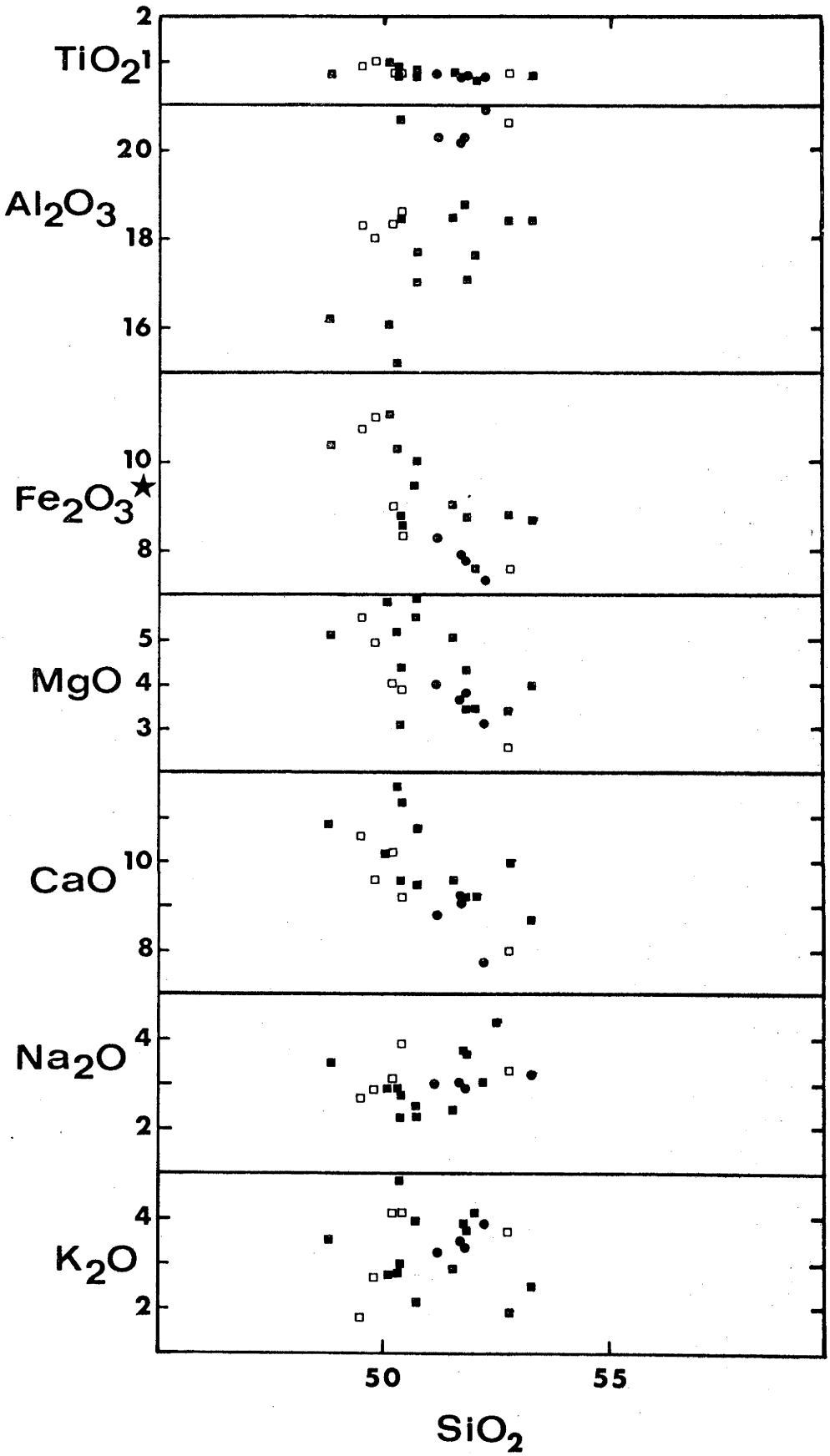

Figure 4.6 Major element oxides of third period samples from the Sabeto area, Viti Levu. Data from Table 4.5, Dickinson et al. (1968), and Rodda (1969). Open squares indicate intrusive monzonite plugs thought to be volcanic centers. Solid squares are samples belonging to the Koroimavua Andesitic Group; solid circles are samples belonging to the Mba Basaltic Group.

island arc suites. Measured Fe_2O_3 contents are high, leading to normative qtz and hyp. When $Fe_2O_3/FeO = 0.25$, most samples are ne-normative.

The most notable features are, of course, the alkali element and other large cation concentrations. K, P, Rb, Ba, Sr, and Pb contents are high and variable. Ti, Zr, Hf, and Nb, which usually are enriched together with these other "incompatible elements", are not. Zr/Hf and Ti/Zr ratios are, however, normal. U and Th are only one to two times more abundant than in second period andesites and Th/U ratios remain about two, but K/U ratios are high (45,000 compared to $\lesssim$10,000 in most crustal materals: see Heier and Rogers, 1963). Average K/Rb and K/Sr ratios are 490 and 22, respectively, and are also similar to those of second period andesites; the Ba/Rb ratio is two times higher. Heat production would be about 1.6 μcal/gm yr.

REE data are given in Table 4.9 and Figure 4.3c for sample 861; poorer quality data for sample 64 appeared earlier (Gill, 1970). Both patterns are inflected with La/Yb ratios of 8.3 and 5.7.

C. Tavua center

Forty years of mining the Au-Ag telluride and sulfide ores of the Tavua Goldfield has resulted in more detailed descriptions of the geology of this area than any other in Fiji (see especially Blatchford, 1955; Cohen, 1962; Denholm, 1966, 1967; and Ibbotson, 1967). These authors describe the development of a large stratovolcano initially made of and built upon an at least 2.5 km thick platform of "olivine basalt" and, later, "trachybasalt". (The difference between these rock types seems to be the presence [in "trachybasalt"] or absence of plagioclase phenocrysts.) A complex period of caldera formation followed and was accompanied by extrusion and intrusion of 1.5 to 2 km of tuffs, breccias, and "andesite" or, more precisely, "biotite andesite", "augite andesite", and/or "trachyandesite", mafic porphyries, and monzonite sills. These "andesitic" and monzonitic magmas were best preserved as dikes and cone sheets encountered more frequently during subsurface mining than in outcrop. Epithermal mineralization apparently post-dated the magmatism.

As the rock names above indicate, Tavua volcanics have traditionally been thought to comprise an oceanic basalt-trachyte association. The designation "andesite", common throughout the literature for the area, apparently derives from the composition of

plagioclase phenocrysts and color indices which are lower than those of the surrounding basalts. Chemically, however, the rocks are a shoshonitic association as recognized by Dickinson et al. (1968), Gill (1970), and Rodda (written comm., 1970), including absarokites, shoshonites, and banakites (following my nomenclature conventions). Petrographically the samples I analyzed are very similar to those described by Ibbotson (1967). Augite phenocrysts predominate, joined by olivine in basic flows and porphyries and labradorite to andesine in less basic ones. Biotite occurs in basic porphyries, monzonites, and the last-erupted differentiates. K-feldspar occurs in monzonites and probably in the groundmass of flows where it is difficult to determine optically. A few per cent oxide phenocrysts are common throughout the suite. The groundmass typically contains pyroxene, feldspar, oxides, glass, and sometimes apatite: zeolites and calcite occur but less commonly than at the Sabeto center.

There is an apparent discrepancy between optical data for pyroxene phenocrysts from the Tavua and Saveto centers which bears on genetic interpretations of these rocks (chapter 12). Ibbotson (1967) reports that augites throughout the Tavua suite have β refractive indices between 1.70 and 1.71 and $2V_\gamma$ angles of 53 to 58°, indicating compositions of about $Ca_{44}Mg_{28}Fe_{28}$, which are curiously Fe-rich for rocks with these bulk compositions. Dickinson et al. (1968) reported similar optic angles for pyroxenes from the Sabeto area but refraction indices $\sim$1.65-1.70, indicating more magnesian compositions. The latter are more like those found in the Wyoming shoshonite province (Nicholls and Carmichael, 1969) or in Taveuni (section 7.1).

In addition to my twelve analyses presented in Table 4.6, Ibbotson (1967) reports twelve others. I shall draw upon his major but not trace element data in the following summary. The Tavua suite provides much greater variety than the Sabeto. However, inspection of Tables 4.5 and 4.6 (also see the summary in Table 8.2) indicates that rocks very similar in composition to Sabeto samples also occur within the Tavua suite (note my samples 897, 898, 918, 896, 888, and 73, and Ibbotson's samples 515, 504, and 505) where they are called "trachybasalts" (504 and 505) or "augite trachyandesite" (515). Apparently, therefore, rocks exposed in the Sabeto area are but one portion of a more complete suite which is better developed (or exposed) at Tavua.

TABLE 4.6. THIRD PERIOD VITI LEVU VOLCANICS: TAVUA AREA.

DATA	893	919	895	72	897	898	918	896A	888	73	76	894
SIO2	48.25	48.37	48.94	49.02	49.46	49.71	51.36	51.66	52.47	53.07	53.66	56.59
TIO2	0.53	0.70	0.57	0.59	0.75	0.68	0.69	0.87	0.66	0.73	0.55	0.56
AL203	12.49	12.17	12.59	12.62	16.57	17.19	18.15	16.46	18.94	17.50	20.86	19.66
FE203	10.32	11.78	10.72	4.96	10.61	10.22	9.18	10.74	8.49	7.06	4.23	6.07
FEO	0.0	0.0	0.0	5.40	0.0	0.0	0.0	0.0	0.0	2.04	2.25	0.0
MNO	0.19	0.20	0.18	0.20	0.24	0.20	0.21	0.20	0.21	0.17	0.16	0.14
MGO	13.12	9.42	10.69	10.67	5.81	4.30	4.01	4.94	3.42	3.72	2.13	1.94
CAO	11.10	12.56	12.03	11.81	9.92	10.10	8.25	9.54	7.78	7.76	8.04	4.98
NA20	1.40	2.00	2.00	2.62	2.22	2.75	2.95	2.80	4.44	3.46	3.45	3.57
K20	2.28	2.45	1.94	1.77	3.73	4.23	4.53	2.43	2.95	3.93	4.15	6.06
P205	0.33	0.36	0.33	0.34	0.69	0.61	0.68	0.36	0.63	0.56	0.52	0.44
(LOI)	3.49	3.09	1.86	3.02	5.97	9.40	3.53	1.54	0.0	1.96	2.95	2.68
(TOTAL)	100.39	99.73	99.59	101.70	98.90	99.18	99.19	99.44	96.11	101.16	99.08	98.87
NA20/K20	0.6	0.8	1.0	1.5	0.6	0.6	0.7	1.2	1.5	0.9	0.8	0.6
MG/MG+FE2	0.76	0.66	0.71	0.70	0.57	0.50	0.51	0.53	0.49	0.49	0.43	0.44
%K	1.89	2.04	1.61	1.47	3.10	3.51	3.76	2.02	2.45	3.26	3.45	5.03
RB	43.	48.	57.	61.	72.	74.	88.	45.	69.	76.	190.	170.
SR	862.	981.	1369.	985.	1370.	799.	1659.	747.	1559.	1053.	2091.	1320.
BA	578.	414.	1361.	693.	968.	835.	1045.	398.	1165.	690.	1030.	457.
PB	5.	5.	6.	9.	11.	7.	10.	6.	11.	10.	14.	16.
K/RB	437.	428.	282.	242.	430.	478.	428.	449.	357.	432.	177.	302.
K/SR	22.	21.	12.	15.	23.	44.	23.	27.	16.	31.	16.	38.
BA/RB	13.3	8.7	23.7	11.4	13.4	11.4	11.9	8.9	17.0	9.1	5.3	2.7
RB/SR	0.050	0.049	0.042	0.062	0.053	0.092	0.053	0.060	0.044	0.072	0.093	0.126
SR87/86	.0	.0	.0	.7039	.0	.7037	.0	.0	.0	.7038	.7039	.7039
TH	.85	.66	.67	.80	1.7	1.0	1.6	.53	2.0	1.1	1.7	3.5
U	.36	.32	.32	.36	.68	.45	.96	.21	.71	.48	.71	1.7
TH/U	2.3	2.1	2.1	2.2	2.5	2.3	1.6	2.5	2.8	2.3	2.4	2.1
K/U	52417.	63465.	50879.	40354.	45723.	78862.	39070.	94081.	34670.	67448.	48709.	30248.
HPU	0.9	0.9	0.8	0.8	1.7	1.5	2.0	0.8	1.6	1.5	1.8	3.3
LA	0.0	0.0	4.8	0.0	14.5	0.0	0.0	0.0	0.0	0.0	0.0	16.5
YB	.0	.0	1.2	.0	2.0	.0	.0	.0	.0	.0	.0	2.2
LA/YB	0.0	0.0	3.9	0.0	7.1	0.0	0.0	0.0	0.0	0.0	0.0	7.6
Y	14.	21.	16.	14.	26.	20.	25.	27.	25.	26.	27.	27.
NI	211.	77.	268.	93.	22.	13.	13.	26.	9.	7.	4.	ND
CO	55.	46.	38.	48.	38.	33.	29.	39.	27.	23.	18.	20.
CR	642.	290.	363.	365.	45.	4.	33.	53.	14.	9.	2.	3.
SC	43.	47.	42.	35.	32.	20.	20.	37.	18.	18.	8.	10.
V	234.	303.	232.	679.	322.	303.	255.	256.	223.	202.	187.	180.
CU	135.	119.	114.	122.	286.	219.	251.	164.	216.	151.	156.	99.
NI/CO	3.8	1.7	7.1	2.0	0.6	0.4	0.4	0.7	0.3	0.3	0.2	0.0
V/NI	1.	4.	1.	7.	15.	23.	20.	10.	24.	29.	45.	0.
MG/NI	376.	742.	241.	690.	1629.	1940.	1930.	1168.	2205.	3180.	3091.	0.
SN	.0	.0	.29	.0	.46	.0	.0	.0	.0	.0	.0	.54
HF	.0	.0	.61	.0	1.2	.0	.0	.0	.0	.0	.0	1.8
ZR	26.	27.	20.	34.	45.	36.	43.	48.	56.	72.	74.	97.
B	0.	0.	10.	0.	0.	0.	10.	8.	12.	0.	0.	0.
ZR/HF	0.	0.	33.	0.	38.	0.	0.	0.	0.	0.	0.	55.
TI/ZR	122.	157.	168.	102.	100.	114.	97.	108.	70.	61.	45.	35.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

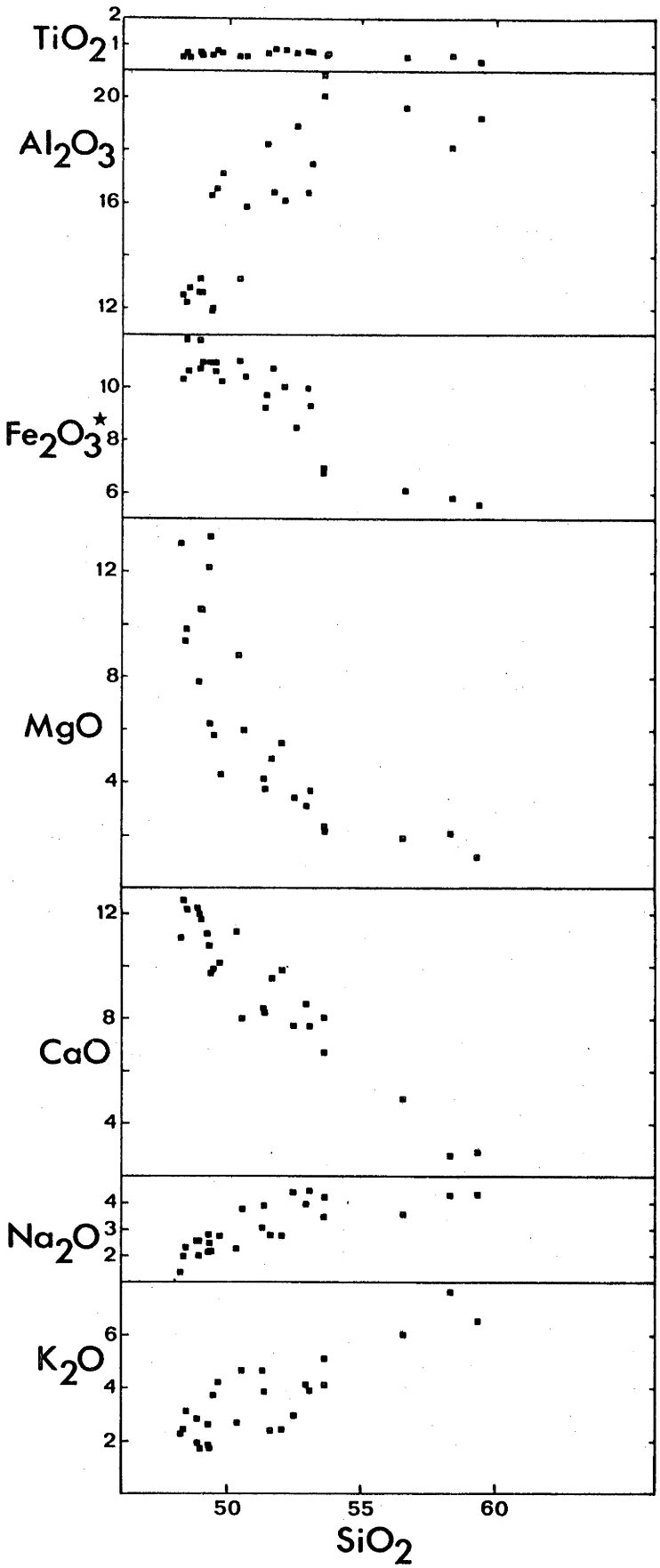

Figure 4.7 Major element oxides of third period samples from the Tavua area, Viti Levu. Data from Table 4.6 and Ibbotson (1967).

Samples 893 and 919 are definitely cumulates of olivine and clinopyroxene and were not completely liquid; others may share a similar history less obviously. There is a wide range of most oxides within the suite, Ti and P varying least. In contrast to the Sabeto area, K_2O increases three to four times with increasing SiO_2 at Tavua. However, if only rocks with 49.5 to 53% SiO_2 are compared, the K_2O-SiO_2 scatter is identical. At both sites average K_2O/Na_2O ratios are <1.0 and average TiO_2 contents are 0.6 to 0.7% -- obvious differences from oceanic alkali basalt-trachyte suites, although most rocks (including differentiates) have normative ne or <2% hyp. Measured Fe_2O_3 contents and LOI are high in both regions.

Large cations, especially Rb and Ba, were also concentrated in Tavua magmas. K/Sr and Ba/Rb ratios are comparable to those of Sabeto samples but K/Rb ratios are lower and Rb/Sr higher. The 87Sr/86Sr ratios are similar. Th and U contents are slightly lower and K/U ratios (averaging 59,000) higher.

REE data for samples 895, 897, and 894 are given in Table 4.9 and Figure 4.3c. Absarokite 895 is notable for its low REE contents, La/Yb ratio of 3.8, and for having only six to seven times chondritic concentrations of heavy REE. Sample 897, with higher Al_2O_3 and lower MgO than 895, has a REE pattern similar to that of sample 861 from the Sabeto area with which it agrees closely in major element composition. Banakite 894, with much higher concentrations of some large cations, has only marginally more La, less of elements Nd to Dy, and similar Y and heavy REE contents relative to the absarokites. K/La ratios are higher than in the high-K calc-alkaline sample from Kadavu.

Ferromagnesian trace element contents have a much wider range but are similar to those of Sabeto samples when rocks with comparable MgO and SiO_2 contents are compared.

Telluride mineralization at Tavua has been thought due to residual fluids from a fractionating magma of which biotite-bearing rocks like sample 894 were the last silicate expression. Usually Sb, Sn, and Cu are also concentrated during Au-Ag-Te mineralization (Vlasov, 1966) and Ni contents of 500 ppm are reported in Tavua sylvanite (Stillwell, 1949). Samples 895, 897, and 894 were therefore mass-spectrometrically analyzed for Au197, Sb123, and Te128 in addition to those elements listed in Table 4.6. Sn and Sb contents increase twofold over this

range, Te contents are erratic, and Cu and Ni decrease. Au was generally undetectable, i.e. less than 0.1 ppm, but a small amount was irregularly dispersed in sample 897.

Tavua Au-Ag mineralization is similar to that of the famous Cripple Creek, Colorado district (Markham, 1960). Its geological environment is also. Lindgren and Ransome (1906, esp. pp. 103-113) provide descriptions and analyses of rocks from Cripple Creek volcano and Lipman et al. (1971) place the volcanism within a regional tectonic context. Both the Tavua and Cripple Creek areas were high-alkali island arc provinces, formed near the end of a phase of volcanic activity. However, the Cripple Creek volcano is 73% "phonolite" with high SiO_2 (>57%), Na>K, and Na_2O+K_2O ∿15%; mafic rocks are uncommon there and in other similar suites in the western interior U.S. (Lipman, 1970).

D. Rakiraki center

The northeastern corner of Viti Levu in the area of the Kauvadra Range has been interpreted as another volcanic center, younger than Tavua. The principal evidence has been a change in dip of lava flows and the ubiquity of dikes and plugs in the area (Hirst, 1965; Seeley and Searle, 1970; Rodda, pers. comm., 1970), to which I add a difference from other centers in magma composition. The most important rock unit of the area is the Nakorotumbu Basalt (Hirst, 1965). These basalts, like the Tavua volcanics, have large, zoned augite phenocrysts with or without olivine and calcic plagioclase. Olivine-bearing rocks predominate in stratigraphically lower sections where plagioclase, if occuring at all, is bytownite. Locally the Nakorotumbu Basalt grades into Tova Andesite, rocks of which are distinguished from basalt only by having less calcic plagioclase and fewer pyroxene phenocrysts (Hirst, 1965).

Five new analyses of Nakorotumbu Basalt are given in Table 4.7 together with one analysis from Foye (1918) and two from Seeley and Searle (1970). There is also one analysis of Tova Andesite (sample 899).

Most oxides show wide variation within the suite; K_2O and TiO_2 the least, averaging 1.8 and 0.8% respectively. The basalts are critically undersaturated with a few per cent normative ne or hyp, depending on the Fe_2O_3/FeO ratio. They are similar in most major element oxide contents to the Tavua and Sabeto suites but differ in minor and trace elements. K and P contents are consistently lower and Na/K ratios higher. Rb, Ba, and Sr contents are somewhat lower,

TABLE 4.7. THIRD PERIOD VITI LEVU VOLCANICS: RAKIRAKI AREA AND NAKOMBALEVU.

DATA	902	AS	2410	2408	912	900	74	899	911	917	916
SIO2	47.21	48.38	48.50	48.52	48.66	49.25	50.14	50.36	52.75	49.97	49.99
TIO2	0.62	0.87	1.23	0.61	1.15	0.70	0.68	0.81	0.76	0.76	0.74
AL203	10.60	19.80	17.56	16.55	16.47	13.66	12.27	15.41	19.00	18.20	17.98
FE203	11.28	3.68	2.35	3.35	10.83	9.67	4.53	9.97	9.28	9.80	10.03
FEO	0.0	6.26	5.11	5.08	0.0	0.0	4.92	0.0	0.0	0.0	0.0
MNO	0.19	0.11	0.14	0.14	0.20	0.17	0.18	0.18	0.20	0.17	0.20
MGO	13.03	5.42	7.56	7.71	7.29	10.70	10.68	7.33	3.97	6.51	6.51
CAO	12.24	10.77	13.38	13.81	10.49	11.59	11.76	10.62	9.70	10.95	11.03
NA20	2.26	2.75	2.35	2.54	2.77	2.10	2.89	2.46	2.97	2.52	2.44
K20	2.09	1.78	1.63	1.62	1.70	1.82	1.56	2.47	1.18	0.89	0.86
P205	0.49	0.19	0.20	0.07	0.43	0.35	0.39	0.39	0.20	0.23	0.21
(LOI)	2.86	2.20	2.40	1.80	1.74	1.93	3.02	1.62	0.32	1.18	1.23
(TOTAL)	99.13	100.10	100.34	100.31	99.35	99.17	100.34	99.18	99.38	98.75	99.53
NA20/K20	1.1	1.5	1.4	1.6	1.6	1.2	1.8	1.0	2.5	2.8	2.8
MG/MG+FE2	0.74	0.55	0.70	0.68	0.62	0.73	0.72	0.64	0.51	0.62	0.61
%K	1.73	1.48	1.36	1.35	1.41	1.51	1.30	2.05	0.98	0.74	0.72
RB	48.	.0	.0	.0	33.	49.	23.	49.	21.	14.	12.
SR	1073.	0.	0.	0.	800.	832.	791.	905.	455.	509.	487.
BA	540.	0.	0.	0.	410.	483.	637.	687.	646.	154.	153.
PB	8.	0.	0.	0.	3.	5.	7.	5.	2.	0.	1.
K/RB	363.	0.	0.	0.	431.	306.	574.	417.	463.	516.	588.
K/SR	16.	0.	0.	0.	18.	18.	16.	23.	22.	15.	15.
BA/RB	11.3	0.0	0.0	0.0	12.5	9.8	28.2	14.0	30.5	10.7	12.5
RB/SR	0.045	0.0	0.0	0.0	0.041	0.059	0.029	0.054	0.047	0.028	0.025
SR87/86	.0	.0	.0	.0	.0	.0	.7041	.7038	.0	.7032	.0
TH	1.4	.0	.0	.0	1.1	.88	1.2	1.7	.58	.84	.75
U	.50	.0	.0	.0	.43	.31	.42	.50	.24	.37	.39
TH/U	2.7	0.0	0.0	0.0	2.6	2.9	2.9	3.5	2.4	2.3	1.9
K/U	34762.	0.	0.	0.	32810.	48978.	30776.	40829.	40469.	20061.	18569.
HPU	1.1	0.0	0.0	0.0	0.9	0.8	0.9	1.3	0.6	0.6	0.6
LA	0.0	0.0	0.0	0.0	0.0	0.0	14.4	0.0	0.0	0.0	0.0
YB	.0	.0	.0	.0	.0	.0	1.0	.0	.0	.0	.0
LA/YB	0.0	0.0	0.0	0.0	0.0	0.0	14.0	0.0	0.0	0.0	0.0
Y	17.	0.	0.	0.	32.	19.	14.	22.	24.	19.	19.
NI	221.	0.	0.	0.	74.	154.	113.	59.	11.	17.	27.
CO	59.	0.	0.	0.	39.	43.	42.	39.	31.	39.	37.
CR	548.	0.	0.	0.	148.	397.	226.	161.	6.	81.	82.
SC	47.	0.	0.	0.	41.	41.	35.	41.	23.	37.	39.
V	239.	0.	0.	0.	307.	237.	247.	284.	230.	266.	238.
CU	122.	0.	0.	0.	152.	115.	94.	176.	91.	61.	137.
NI/CO	3.7	0.0	0.0	0.0	1.9	3.6	2.7	1.5	0.4	0.4	0.8
V/NI	1.	0.	0.	0.	4.	2.	2.	5.	21.	15.	9.
MG/NI	355.	0.	0.	0.	596.	418.	570.	744.	2155.	2253.	1430.
SN	.0	.0	.0	.0	.0	.0	1.0	.0	.0	.0	.0
HF	.0	.0	.0	.0	.0	.0	.82	.0	.0	.0	.0
ZR	44.	0.	0.	0.	87.	50.	68.	53.	65.	64.	59.
B	0.	0.	0.	0.	0.	0.	0.	11.	28.	9.	9.
ZR/HF	0.	0.	0.	0.	0.	0.	83.	0.	0.	0.	0.
TI/ZR	86.	0.	0.	0.	79.	83.	60.	91.	70.	72.	75.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

AS FROM FOYE(1918); 241C AND 2408 FROM SEELEY AND SEARLE(1970)

Ba/Rb and Th/U ratios higher, and K/U lower. Two 87Sr/86Sr ratios are 0.7038 and 0.7041. One earlier REE pattern (Gill, 1970, Figure 7) has a higher La/Yb ratio and lower Y and heavy REE contents than do most rocks of the more shoshonitic suites.

The lone analyzed Tova Andesite cannot be a differentiate of Nakorotumbu Basalt even though its gross major element composition is consistent with such an origin. It is quartz-normative, and lower in K, P, Rb, Sr, Pb, Th, and U--all elements concentrated by fractionation processes.

E. Vatia center and other hornblende-bearing third period volcanic rocks

Unlike the three third period centers just discussed, rocks of the Vatia Peninsula are entirely basaltic-andesites and andesites, usually hornblende-bearing. They are described by Ibbotson (1962) who regards their chronologic relationship to neighboring shoshonites as obscure. Rodda et al. (1967) report a single 9.2 ±2 m.y. age for feldspar in a Vatia sample.

Vatia rocks commonly contain coarse hornblende crystals with strong reaction rims and some alteration to chlorite, calcite, and an opaque. Phenocrystic plagioclase is zoned from labradorite to andesine; normative plagioclase is labradorite. Augite often accompanies hornblende but also occurs without it; several per cent magnetite is common. The groundmass is a fine-grained composite of pyroxene, feldspar, an opaque, and glass.

Analyses of four samples from the peninsula are given in Table 4.8 in addition to which Ibbotson (in Rodda, 1969, p. 13) provides three. All are very different from other third period volcanics, and even from those with similar SiO_2 contents. They are, instead, similar to the Namosi Andesites. Relative to these they are consistently more aluminous and poorer in Mg, Fe, and Na at comparable SiO_2 levels. $Mg/Mg+Fe^{+2}$ ratios are similar but Na_2O/K_2O ratios are lower. K, Rb, Ba, Sr, and Pb contents are higher in Vatia than Namosi rocks but not uniformly so: Rb, Ba, and Sr are slightly enriched relative to K in the Vatia samples. One 87Sr/86Sr ratio measurement (0.7043) is higher than Namosi analyses. Th and U contents in Vatia samples are only half those of Namosi ones, resulting in similar Th/U ratios but K/U ratios (about 40,000) more akin to other third period suites. Ti and especially Zr contents are also lower, resulting in higher Ti/Zr ratios.

TABLE 4.8. THIRD PERIOD VITI LEVU VOLCANICS: VATIA PENINSULA.

DATA	891	889	890	892	885
SIO2	52.53	56.71	58.50	58.75	55.45
TIO2	0.78	0.68	0.57	0.66	0.59
AL203	19.79	18.60	18.22	18.62	19.19
FE203	7.96	7.24	6.36	6.51	7.65
FEO	0.0	0.0	0.0	0.0	0.0
MNO	0.14	0.14	0.15	0.12	0.13
MGO	4.39	3.13	3.47	2.90	3.32
CAO	9.96	8.53	7.66	7.44	8.70
NA20	2.72	3.28	3.69	3.55	2.97
K20	1.46	1.47	1.20	1.28	1.76
P205	0.26	0.23	0.16	0.16	0.23
(LOI)	2.15	1.86	2.06	1.48	2.48
(TOTAL)	99.31	99.42	99.49	99.63	99.12
NA20/K20	1.9	2.2	3.1	2.8	1.7
MG/MG+FE2	0.57	0.51	0.57	0.52	0.51
%K	1.21	1.22	1.00	1.07	1.46
RB	24.	28.	21.	24.	39.
SR	821.	745.	731.	586.	626.
BA	481.	620.	606.	455.	538.
PB	5.	4.	5.	4.	3.
K/RB	513.	440.	486.	436.	371.
K/SR	15.	16.	14.	18.	23.
BA/RB	20.3	22.4	29.5	18.6	13.7
RB/SR	0.029	0.037	0.028	0.042	0.063
SR87/86	.0	.0	.0	.7043	.0
TH	.68	.80	.74	.94	.48
U	.27	.42	.29	.43	.29
TH/U	2.5	1.9	2.6	2.2	1.6
K/U	45338.	28953.	34688.	24904.	50401.
HPU	0.7	0.8	0.6	0.8	0.7
LA	0.0	0.0	0.0	0.0	0.0
YB	.0	.0	.0	.0	.0
LA/YB	0.0	0.0	0.0	0.0	0.0
Y	26.	27.	23.	68.	23.
NI	14.	10.	9.	11.	13.
CO	22.	18.	15.	21.	23.
CR	14.	9.	11.	5.	25.
SC	42.	24.	20.	19.	19.
V	262.	209.	181.	190.	207.
CU	106.	67.	35.	58.	83.
NI/CO	0.7	0.6	0.6	0.5	0.6
V/NI	18.	20.	20.	17.	15.
MG/NI	1840.	1840.	2265.	1563.	1489.
SN	.0	.0	.0	.0	.0
HF	.0	.0	.0	.0	.0
ZR	39.	45.	54.	59.	37.
B	9.	8.	33.	19.	17.
ZR/HF	0.	0.	0.	0.	0.
TI/ZR	120.	90.	63.	67.	95.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

Hornblende-bearing rocks are also known at other third period eruptive centers. As mentioned in section 4.4B, the occurrence of hornblende contributed to the designation of Sabeto center rocks as "andesitic", but no hornblende-bearing samples have been analyzed. Amphibole is uncommon within the Tavua crater area itself (see Ibbotson, 1967, p. 33) but I collected hornblende-bearing sample 885 in a quarry 12 km west of Ba, north of the Sabeto Range, within an area mapped as Mba Basalt (Bartholomew, 1959a). Seeley and Searle (1970) document the widespread occurrence of such rocks near Rakiraki, call them the Narewa Andesites, and provide two analyses. They are also associated with basalts of similar age in the Lomaiviti and Lau island groups (see sections 6.4B and 7.2).

In none of these cases are the hornblende basaltic-andesites and andesites likely differentiates of the shoshonitic or basaltic rocks near which they occur. Seeley and Searle conclude that their Narewa Andesites result from "a mildly alkaline basalt parent following a typically calc-alkaline differentiation trend" (1970, p. 68), involving fractionation of clinopyroxene, plagioclase, and minor magnetite from Nakorotumbu Basalt. Hirst (1965, p. 42) did describe one instance in which pyroxene in Nakorotumbu Basalt apparently reacted with liquid to yield hornblende. However, it is petrologically unlikely that critically undersaturated basalt will fractionate to yield acid andesite at low pressures (Yoder and Tilley, 1962), and the relatively low large cation minor and trace element contents of other hornblende-bearing third period andesites clearly preclude their derivation from Nakorotumbu Basalt or the voluminous shoshonitic magmas further west.

Earlier (Gill, 1970, p. 193) I speculated that these hornblende andesites might have been erupted simultaneously with the chemically and petrographically similar Namosi Andesites, based on the lone K-Ar date mentioned above, the stratigraphic ambiguity, and the similarities in composition. The association of such rocks with other young basaltic centers in Fiji weakens such an argument.

F. Other centers

Ibbotson (1960) and Band (1968) describe the Nakombalevu Basalt, 13 km^2 of which crops out northwest of Suva and which is thought to be about 150 m thick. These basalts are included with the Mba Group because of similarities in petrography and stratigraphic position.

Remarkably similar analyses are given in Table 4.7 for two randomly selected cobbles collected about 2000 m downstream from their outcrop area. This and the petrographic similarity of these samples to those described by Ibbotson and Band suggest that the Nakombalevu Basalt is a rather uniform unit befitting its limited size.

Although modally they are also olivine basalts, these are hypnormative in strong contrast to other third period Viti Levu mafic rocks. Their TiO_2 and major element oxides are similar, but K, P, and the large cation trace elements are much less abundant, and Na/K ratios are higher. At 550, the mean K/Rb ratio is also higher; K/Sr, Rb/Sr, and 87Sr/86Sr ratios are lower, as are Th contents and K/U ratios.

Band (1968) also discusses the shoshonitic Mbengga Volcanic Group which crops out on two islands just off the south coast of Viti Levu. No analyses are available but Band describes a suite containing volcanic rocks rich in olivine, augite, and labradorite $\pm$ hypersthene and resorbed hornblende; another suite without olivine but containing 0-10% hornblende, 3-10% biotite, and apatite; and intrusives consisting of plagioclase rimmed with K-feldspar, augite, hypersthene, and biotite.

G. Summary

There is considerable diversity among what are grouped together as third period volcanic rocks (see Table 8.2). Those of the Sabeto and Tavua centers are shoshonitic. Apart from somewhat higher Ba and Rb and lower U contents at Tavua, Sabeto rocks are analogous in composition to one portion of the Tavua suite and can be considered a specialized case of the latter, more fully developed (or exposed) situation. Large cation trace elements are concentrated in both suites without concomitant enrichments of Ti, Zr, Hf, or Nb. REE patterns are similar for rocks of similar Al_2O_3 and MgO contents, and light REE are less concentrated than in calc-alkaline rocks with comparable alkali contents. Th/U ratios remain low in both areas but K/U ratios are among the highest known.

Most of these characteristics also apply, but to a lesser extent, to the Rakiraki center which illustrates the difficulty of distinguishing between shoshonite and alkali basalt suites. Ti, Zr, Hf, and Nb contents remain low and K, P, Rb, Ba, Sr, and Pb contents are high, but not as high as Sabeto or Tavua rocks so that the average Na_2O/K_2O ratio is 1.4, not 1.0. Rakiraki area basalts are generally

Table 4.9 Rare earth analyses of third period, Viti Levu samples (also see Gill, 1970, Table 3)

	861	895	897	894
La	15.0	4.7	13.5	15.9
Ce	31.0	10.5	22.6	30.2
Pr	5.0	1.9	3.9	4.5
Nd	22.5	8.1	20.0	18.0
Sm	5.6	2.2	4.9	3.8
Eu	1.6	.73	1.4	1.2
Gd	4.9	2.2	4.9	4.1
Tb	.69	.40	.75	.68
Dy	3.7	2.0	-	3.3
Ho	.78	.49	.79	.74
Er	1.8	1.2	1.9	2.0
Yb	1.8	1.2	1.9	2.1

less undersaturated. Apart from being basaltic, the only features shared between the composition of rocks from these three centers and the Nakobalevu center in southeastern Viti Levu is an average TiO_2 content of about 0.7%. This center apparently produced relatively uniform olivine tholeiite with much lower K, P, Rb, Ba, Sr, Pb, Th, and U contents and 87Sr/86Sr ratios. Such tholeiite could be parental to the Tova Andesite sample analyzed, whereas Nakorotumbu Basalt could not.

Finally, there are hornblende-bearing basaltic-andesites and andesites along the northern coast of Viti Levu and at the Beqa center off the south coast. These apparently belong to the third period of volcanism despite the 9 m.y. K-Ar age, but cannot be differentiates of the shoshonitic rocks near which they occur. They are similar to yet different from the composition of Namosi Andesites and the differences (higher Al, K, Rb, Ba, Sr, and Pb, and lower Th, U, Ti, and Zr contents; lower K/Rb and K/Sr and higher 87Sr/86Sr ratios) are in the direction of shoshonitic compositions.

4.5 Plutonism

A. Petrography and age

Plutonics ranging from olivine gabbro to quartz tonalite intrude first period rocks and appear as detritus in basal second period sediments. Their occurrence led Woolnough (1903) to think Fiji

"continental". Similar plutonics intrude first period rocks throughout the Outer Melanesian Arc (Kermadec: Brothers and Searle, 1970; Tonga: Ewart et al., 1972; New Hebrides and Solomons: Coleman, 1970; and New Britain: Macnab, 1970). Yet the precise age and genesis of the Tholo Plutonics on Viti Levu remain ambiguous.

Rodda (1967) describes the geologic setting and earlier (Gill, 1970) I summarized the published petrography of these plutonics. Some rocks appear to record two stages of amphibole growth and/or to have experienced some recrystallization of ferromagnesian minerals in interstices between large feldspar crystals.

McDougall (1963) reported one 33 ±1 m.y. age measurement for hornblende from a western Viti Levu tonalite and Rodda et al. (1967) a 50 ±5 m.y. result for hornblende from a southwestern Viti Levu gabbro. The latter authors also report K-Ar age measurements of 7-13 m.y. for biotite and hornblendes from three tonalites further to the east and closer to the outcrop area of second period volcanics. One of these is from the Wainikoro stock which Band equated with the second period Mau Quarry stock as "Mio-Pliocene quartz diorite" on his map (sheet 19), a decision he later recanted (1968, pp. 29-30). My sample from the Wainikoro stock (870, in Table 4.3) is identical to Namosi Andesite in composition and very different from Tholo plutonics in K, Rb, Ba, Sr, Pb, Rh, and U contents. (Its major elements are intermediate between the bimodal Tholo compositions [Table 4.10] but Rodda [pers. comm., 1971] mentions unpublished analyses of Tholo rocks which also have more intermediate compositions.) Namosi Andesites have not been radiometrically dated, but are believed to be Tertiary g (9 to 5.5 m.y.: Figure 1.4) in age (see section 4.3).

Thus there are three possible interpretations for the age of Tholo plutonism. First, the data of McDougall and Rodda et al. can be accepted (as in Rodda, 1967) as prima-facie evidence for discontinuous plutonism spanning 40 m.y. and being younger in the east. If so the younger plutons must have been uplifted and eroded within a very few million years to appear as detritus in Tertiary g sediments. Second, the ages may reflect time of uplift rather than emplacement. Many plutonics and most volcanics which they intrude contain mineral assemblages indicative of burial metamorphism (see section 4.2) which are usually interpreted as indicating temperatures of 200-400°C and depths of several kilometers (e.g. Herrmann and Wedepohl, 1970).

Hornblende and especially biotite may lose considerable Ar if held at such temperatures for 10^6 years or longer (Dalrymple and Lanphere, 1969, pp. 151-163 and 172-5). Uplift terminating burial may have occurred sooner in the west where topographic relief is less. In the east, 7-13 m.y. would be the approximate time during which an erosional surface developed which later became the unconformable contact between rocks of my first and second periods. Third, the 7-13 m.y. ages may record effects of second period volcanism opening and reclosing plutonic K-Ar systems. These ages are close to that inferred for Namosi volcanism and have been found only in plutons spatially contiguous with such volcanics. The Wainikoro stock, identical in composition to Namosi Andesites but in age to these other Tholo Plutons, suggests this last alternative as does petrographic evidence for two periods of amphibole growth. If so, there are insufficient data with which to judge the primary age of Tholo plutonism.

B. Composition

Thirteen new analyses of plutonic rocks and three of associated dikes are given in Table 4.10. Rodda (1969) lists four additional analyses. Together these form a bimodal population of gabbros (45-51% SiO_2) and tonalites (64-77% SiO_2); the dikes are mafic.

The gabbros appear to be cumulates of plagioclase, clinopyroxene, and magnetite, with or without olivine and amphibole. Layers of plagioclase and pyroxene crystal "sediments" were observed in one outcrop. Compositions are as expected in such cumulates: Ti, Na, K, and P contents are low and Na/K ratios high; similarly, Rb, Ba, Sr, Pb, Th, U, Zr, and Y contents are low, K/Rb ratios high, and K/Sr and Rb/Sr ratios low. None of these elements is accomodated in the lattices of plagioclase or pyroxene and the element ratios are those of plagioclase-dominated cumulates (Gill and Murthy, 1970). The high Al and erratic Mg, Fe, and Ca contents differ from first period basalts and are also consistent with an accumulative origin.

Only the three dike compositions are likely to represent former basic magmas. These are richer in Ti, Zr, Na, Y, and P, and poorer in Al than the gabbros, but similar in most other major and trace elements. They resemble first period basalts except for being richer in Si and ferromagnesian trace elements when rocks of comparable MgO contents are compared. All three rocks are altered and secondary quartz appears to have precipitated in sample 58, making confirmation

TABLE 4.10. THOLO PLUTONICS.

DATA	859	852	845	849	846	847	57	56	58	909	910	908	858	867	848	59
SIO2	45.44	45.86	46.15	46.61	46.68	51.07	51.45	51.52	53.20	63.96	63.70	69.69	70.15	72.53	74.56	76.72
TIO2	0.16	0.41	0.58	0.13	0.42	0.70	1.36	1.45	1.28	0.50	0.47	0.61	0.37	0.31	0.38	0.23
AL203	19.44	27.97	21.14	19.67	23.77	19.29	17.03	16.93	15.76	16.66	17.31	14.77	14.87	14.42	12.85	12.81
FE203	6.40	6.33	11.10	5.62	8.05	9.86	2.97	3.57	3.96	5.89	5.59	4.57	3.57	2.94	3.42	1.43
FEO	0.0	0.0	0.0	0.0	0.0	0.0	6.23	5.91	7.84	0.0	0.0	0.0	0.0	0.0	0.0	0.31
MNO	0.11	0.09	0.19	0.11	0.13	0.17	0.22	0.23	0.20	0.16	0.13	0.11	0.07	0.09	0.09	0.04
MGO	11.39	2.35	6.37	9.99	4.67	4.85	7.22	7.01	4.97	2.13	2.16	0.78	1.16	0.82	0.68	0.44
CAO	16.27	15.54	13.21	17.29	15.03	11.43	10.17	9.95	8.90	6.44	6.62	3.41	3.94	2.90	3.16	2.05
NA20	0.67	1.30	1.13	0.52	1.16	2.29	2.87	2.94	3.43	3.48	3.40	5.06	4.04	4.73	4.48	5.69
K20	0.10	0.14	0.12	0.05	0.07	0.25	0.22	0.20	0.37	0.57	0.49	0.87	1.76	1.20	0.32	0.24
P205	0.02	0.02	0.01	0.01	0.01	0.08	0.28	0.29	0.10	0.19	0.13	0.13	0.08	0.07	0.06	0.04
(LOI)	1.00	0.96	0.14	1.42	0.86	2.83	2.82	2.80	3.55	1.54	0.72	0.74	0.65	1.34	0.66	0.84
(TOTAL)	99.53	99.42	100.61	99.41	99.94	100.01	99.13	99.96	99.19	99.12	99.22	98.86	99.03	99.09	99.78	99.02
NA20/K20	6.6	9.1	9.5	10.2	16.4	9.3	13.1	15.1	9.4	6.1	7.0	5.8	2.3	3.9	13.9	23.3
MG/MG+FE2	0.81	0.47	0.58	0.81	0.58	0.54	0.64	0.63	0.49	0.47	0.48	0.29	0.44	0.40	0.32	0.38
%K	0.08	0.12	0.10	0.04	0.06	0.21	0.18	0.16	0.30	0.48	0.40	0.72	1.46	0.99	0.27	0.20
RB	1.0	1.9	.60	.82	.61	3.1	1.7	1.1	5.1	7.6	7.2	8.7	21.	13.	.91	1.9
SR	239.	195.	161.	90.	174.	162.	262.	259.	171.	360.	338.	150.	207.	193.	111.	125.
BA	20.	26.	26.	10.	22.	35.	89.	92.	46.	179.	162.	274.	315.	340.	91.	43.
PB	0.	0.	0.	0.	0.	0.	0.	0.	3.	0.	0.	2.	0.	2.	0.	0.
K/RB	830.	612.	1660.	519.	968.	664.	1090.	1434.	593.	628.	561.	830.	684.	747.	2952.	1049.
K/SR	4.	6.	6.	5.	3.	13.	7.	6.	18.	13.	12.	48.	70.	51.	24.	16.
BA/RB	20.0	13.7	43.3	12.5	36.7	11.3	53.8	80.9	9.0	23.6	22.5	31.6	14.8	25.5	100.0	22.1
RB/SR	0.004	0.010	0.004	0.009	0.003	0.019	0.006	0.004	0.030	0.021	0.021	0.058	0.103	0.069	0.008	0.015
SR87/86	.0	.0	.0	.0	.0	.0	.7032	.7046	.7033	.0	.0	.0	.0	.0	.0	.7043
TH	ND	ND	ND	ND	ND	ND	.21	.21	ND	.61	.61	.98	1.7	2.3	ND	2.8
U	ND	ND	ND	ND	ND	ND	.21	.10	ND	.23	.20	.38	.79	.55	ND	1.1
TH/U	0.0	0.0	0.0	0.0	0.0	0.0	1.0	2.0	0.0	2.7	3.0	2.6	2.2	4.1	0.0	2.5
K/U	0.	0.	0.	0.	0.	0.	8716.	15772.	0.	21131.	19923.	19071.	18412.	17986.	0.	1811.
HPU	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0.2	0.0	0.4	0.4	0.7	1.3	1.1	0.0	1.4
LA	0.0	0.0	0.0	0.0	0.0	0.0	8.8	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	11.2
YB	.0	.0	.0	.0	.0	.0	3.0	.0	.0	.0	.0	.0	.0	.0	.0	2.3
LA/YB	0.0	0.0	0.0	0.0	0.0	0.0	2.9	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	4.8
Y	4.	6.	7.	4.	7.	24.	29.	31.	27.	23.	20.	56.	22.	27.	55.	17.
NI	61.	4.	15.	59.	24.	28.	85.	74.	17.	ND	4.	ND	ND	ND	ND	3.
CO	50.	21.	45.	37.	31.	34.	34.	31.	33.	12.	13.	5.	6.	7.	3.	4.
CR	195.	5.	36.	798.	30.	46.	177.	175.	13.	6.	8.	5.	4.	8.	3.	2.
SC	16.	17.	37.	42.	32.	31.	27.	28.	28.	10.	13.	15.	9.	5.	14.	4.
V	101.	231.	440.	150.	375.	254.	218.	237.	272.	48.	83.	15.	64.	31.	22.	26.
CU	10.	74.	89.	19.	100.	80.	40.	49.	146.	4.	51.	8.	8.	3.	3.	1.
NI/CO	1.2	0.2	0.3	1.6	0.8	0.8	2.5	2.4	0.5	0.0	0.3	0.0	0.0	0.0	0.0	0.8
V/NI	2.	57.	29.	3.	16.	9.	3.	3.	16.	0.	20.	0.	0.	0.	0.	9.
MG/NI	1128.	3483.	2574.	1018.	1164.	1052.	511.	570.	1791.	0.	3212.	0.	0.	0.	0.	865.
SN	.0	.0	.0	.0	.0	.0	2.0	.0	.0	.0	.0	.0	.0	.0	.0	2.1
HF	.0	.0	.0	.0	.0	.0	1.8	.0	.0	.0	.0	.0	.0	.0	.0	3.9
ZR	5.	8.	10.	4.	7.	38.	126.	132.	75.	87.	76.	190.	104.	118.	77.	124.
B	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.
ZR/HF	0.	0.	0.	0.	0.	0.	71.	0.	0.	0.	0.	0.	0.	0.	0.	32.
TI/ZR	192.	300.	348.	195.	360.	110.	65.	66.	102.	35.	37.	19.	21.	16.	30.	11.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

of their relationship to the volcanic rocks difficult. REE contents should remain unaffected by this degree of alteration (see section 4.2D) but are strikingly different from those of first period volcanics (Table 1b in Gill, 1970, and Figure 4.3d).

Tonalite analyses may reflect magma compositions. None available to me are intermediate; all are plutonic equivalents of dacitic or rhyolitic extrusives and almost identical in composition to such rocks of the first period suite (compare Tables 4.1 and 4.10). The plutonics have slightly lower Ti, Zr, Y, and Na contents. As in the volcanic suite there are a few samples with higher alkali contents than the others. Rb and Ba contents are comparable or slightly lower in the plutonic rocks whereas Sr and the ferromagnesian trace element contents are consistently higher. Th and U contents are higher in the plutonics which may be a primary feature or may reflect less alteration and/or greater difficulty removing these elements from coarse crystals than a volcanic groundmass. Th/U ratios are higher than observed in the volcanics. REE contents of the most SiO_2-rich tonalite (sample 59; Figure 4.3d) are similar to those of the basic dike (sample 57) and very dissimilar to those of first period volcanics, especially the rhyolite (Figure 4.3a).

No 87Sr/86Sr ratios other than those reported earlier (Gill, 1970) were measured. The isotopic discrepancy between otherwise isochemical samples 56 and 57 has no analytical explanation and is ascribed to addition of more radiogenic Sr to sample 56 during alteration. If so, the remaining analyses conform to the positive correlation between Rb/Sr and 87Sr/86Sr ratios observed in first period volcanics (see section 4.2C and Gill and Compston, Appendix 3).

C. Relationship to volcanism

The voluminous batholiths of orogenic regions are among the most studied of all igneous rocks and are often considered to be remnants of magma chambers that once fed volcanoes with which they are thought to have been associated in time and space (i.e. Hamilton and Myers, 1967; Hamilton, 1969; or Dickinson, 1970a,b). In some cases, variations in their chemical compositions are used to infer subduction polarity (Bateman and Dodge, 1970; Dickinson, 1970a; Macnab, 1970), just as with volcanic rocks (see chapter 3). Tholo plutonism should, therefore, be genetically related to some period of Viti Levu volcanism.

Houtz (1960), Dickinson (1967), and Gill (1970) used the 33 and 50 m.y. age determinations, geographic positions, stratigraphic relationships, and chemical composition of Tholo plutonics to infer simultaneity with first period volcanism. Tholo plutons would have been roots of Wainimala volcanoes, or would at least have been related magmas which crystallized without reaching the surface. They could thus span the age range of such volcanism. Because the 7-13 m.y. ages of some plutons are younger than the youngest faunas known to be associated with first period volcanics, either these ages do not reflect times of crystallization (see section 4.5A) or some Tholo plutons are related to the first and others to the second period of volcanism. Data available to me are insufficient to assess this but analyses in Table 4.10 and Rodda (1969) suggest no change in composition with time. Moreover, rocks with the large cation contents of Tholo tonalites cannot be differentiates of Namosi Andesites. I therefore regard the 7-13 m.y. ages as secondary.

The bimodal population of Tholo pluton compositions and their REE and ferromagnesian trace element contents are, however, inconsistent with them being simple analogues of first period volcanism. A possible explanation would be to regard the gabbros as cumulates and the much more voluminous tonalites as remelted first period volcanics and volcaniclastics. Many studies show that the low-melting fraction of such rocks at about 25 km depths will be granitic to granodioritic in composition (see Robertson and Wyllie, 1971, and references therein). Low K_2O in first period rocks would preclude significant orthoclase in their subsolidus assemblages and make possible quartz tonalite upon initial melting or tonalite with increased melting. Depending on H_2O and CO_2 contents, there can be a wide temperature interval in which silica-rich magma co-exists with calcic plagioclase and either pyroxene or biotite and hornblende. These circumstances might yield magmas of similar major element composition to first period rhyolites and dacites but with higher water contents, making them incapable of sufficient vertical movement to reach the surface (Burnham, 1967). Quartz and plagioclase would be the or among the first phases to melt in an orthoclase-free assemblage and might yield melts with inflected REE patterns and more Sr than K, Rb, or Ba.

This is not a satisfactory alternative as it leaves unexplained the inflected REE pattern of the basic dike, and the similarities in

alkali yet dissimilarities in ferromagnesian trace element contents between plutonics and volcanics. It would also predict Eu-enrichment in the tonalite (which was not observed) unless the feldspar being melted was very anorthite-rich (Philpotts, 1970). I have no other explanation but will return to the question of crustal anatexis in chapters 9 and 13.

Chapter 5. Geology and Geochemistry of Vanua Levu

5.1 Introduction

Vanua Levu is the second largest Fiji island (10,000 km^2), elongate in shape, and on the average about 30 km in width. The strike of elongation apparently reflects a former alignment of volcanic centers. It is separated from Viti Levu by 65 km of shallow water (<1000 fathoms) known as the Bligh Water in memory of the Bounty's captain who sailed through it in 1789 after the mutiny, pursued by two Fijian war canoes.

The geological history of Vanua Levu is less well known than that of Viti Levu but, perhaps as a consequence, appears simpler. A summary based on reconnaissance mapping is given by Rickard (1966); 1:50,000 scale mapping is partially complete (see Bartholomew, 1959b; Ibbotson, 1969; Rickard, 1970; Coulson, 1971; and brief resumés in the annual reports of the FGSD for 1967-69). In general, there is evidence of two distinctly different periods of volcanism. The first is preserved as the predominantly intermediate and submarine rocks of the Mathuandrove Super-Group (see definition below), a largely island arc tholeiitic series of Late Miocene to Early Pliocene age. The second volcanic episode occured mainly in the southwest and yielded a largely subaerial basalt suite, the younger Mbua Basaltic Group.

The older rocks form the island's exposed basement but probably overlie strata similar in age and origin to the basement rocks of Viti Levu (see section 4.2A). Both islands appear to be emergent portions of the Lau-Colville Ridge and are likely to have had a common origin and initial history. They have comparable Bouguer anomalies (J. Worthington, pers. comm., 1971) and horizons thought equivalent to first period Viti Levu rocks have been followed across the shallow Bligh Water at depths of several kilometers during recent geophysical explorations for oil.

Locations of the 24 Vanua Levu samples analyzed are indicated in Figure 5.1 and listed together with petrographic descriptions in Appendix 2.

5.2 Mathuandrove Super-Group volcanism: Late Miocene to Early Pliocene

A. Age and environment

Because I wish to consider only the largest stratigraphic units having petrogenetic significance, and to avoid local problems of

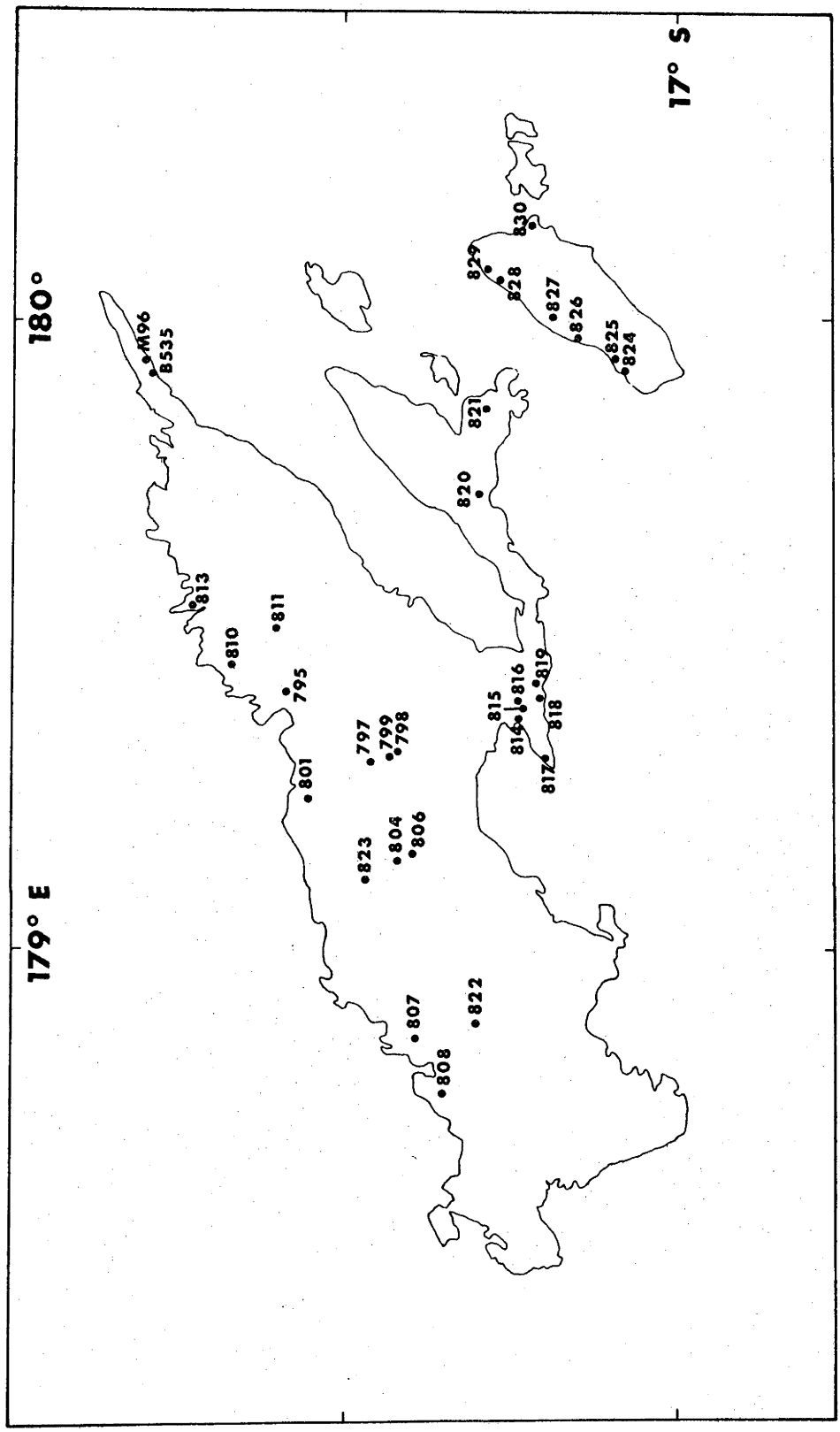

Figure 5.1 Location of Vanua Levu and Taveuni samples

nomenclature, I shall follow the suggestion of W. Hindle by using the name Mathuandrove Super-Group (MSG) to denote rocks of the Natewa, Naroro, Monkey Face, and Undu Groups which were defined by Rickard (1966) and Coulson (1971). Interbedded sediments contain some faunas Tertiary g and h or Neogene Zones 15 to 20 in age (Rickard, 1966; Blow, 1969, Figure 37). (A K-Ar dating project by I. McDougall and W. Hindle is to begin soon but results are not yet available). Volcanism producing rocks of this group is thought to be Late Miocene to Early Pliocene on available evidence, and therefore to be contemporaneous with the second period of Viti Levu volcanism.

Most volcanic activity took place in a shallow marine environment. Vanua Levu apparently represents the coalescence of several volcanoes which began with submarine activity above a pre-existing ridge and ended as subaerial cones with intervening marine and subaerial basins. Most of the island has been uplifted considerably since at least the Pliocene. Sample 798 is the core of a lava pillow collected at an elevation of about 935 m.

B. Petrography

This summary is based on descriptions by Rickard (1966, 1970), Bartholomew (1959b), Ibbotson (1969), and Coulson (1971) as well as my own observations.

MSG rocks are petrographically diverse. Dacites and rhyolites of Rickard's Undu Group are spatially distinct and will be considered separately, but the majority are described as andesites, basaltic-andesites, and basalts, with basaltic-andesites predominating. Virtually all have phenocrysts of plagioclase, ranging from bytownite to andesine, and most contain augite and minor oxides. Hypersthene phenocrysts commonly accompany and may be more or less abundant than augite and seem most common in acid andesites (>60% SiO_2). Hornblende-bearing volcanics are uncommon but known (e.g. samples 815 and 820). Intrusive late-stage andesitic plugs contain up to 15% hornblende phenocrysts along with plagioclase, augite, hypersthene, and magnetite (e.g. samples 804 and 822). Olivine-bearing basalts occur locally in the west (Coulson, 1971), near Savusavu (samples 814 and 818), and on Cakaudrove Peninsula (sample 821). Matrices usually contain phenocryst minerals and are glassy to holocrystalline; trachytic textures occur but are not common. Pigeonite has not been found. Some

rocks have been altered so as to include chlorite, calcite, and zeolites, but not on a regional scale.

The dacites and rhyolites occur chiefly on Udu Peninsula. They are thought to be Upper Pliocene in age, based on foram faunas and stratigraphic correlations, and therefore the youngest volcanic rocks of the MSG (Rickard, 1970; Ibbotson, 1969). The absence of local negative Bouguer anomalies (J. Worthington, pers. comm., 1971) indicate these light rocks are only a thin veneer. They are mostly laminated tuffs, breccias, and flows of diverse variegation, containing phenocrysts of andesine, quartz, an oxide, and traces of hornblende, hypersthene, and clinopyroxene. Modal alkali feldspar, probably sodic, and labradorite have been observed. Rickard (1970, p. 5) suggests that the diversity of these rocks represents effects of cooling ignimbritic ash flows.

C. Geochemistry

Twenty new analyses of MSG rocks are given in Tables 5.1 and 5.2 together with one from Jakes (1970) and another from Hindle (1970) which I analyzed for trace elements only. Hindle analyzed about twenty more MSG samples, Rodda (1969) compiled results for an additional fifteen, and there were at least eight unpublished analyses by FGSD staff by mid-1971. These major element data are summarized in Figures 5.2 and 5.3.

They, like analyses of first period Viti Levu rocks, define a suite with wide silica range, around one per cent TiO_2, moderate to high Al_2O_3, variable but often significant Fe and Ti enrichment, and somewhat low but variable K_2O and P_2O_5. In analyses from Hindle and Rodda Fe_2O_3/FeO ratios are 0.6 ±0.2, but even if it is assumed to be 0.25, $Mg/Mg+Fe^{+2}$ ratios in basalts can exceed the 0.65 minimum for melts in equilibrium with mantle olivine. Only low-pressure hornblende gabbro xenoliths have been found, however (e.g. Rickard, 1966).

To make sense of a reconnaissance study such as this, one hopes for spatial and temporal regularity in processes of magma generation. Such expectation is partially vitiated in this case so that the available analyses and stratigraphic control are inadequate to explain fully the variability observed. Some is due to crystal accumulation and other, especially in Na, Ca, and Sr contents, witness alteration effects. But some regional patterns are also apparent.

The majority of basaltic-andesites and andesites, which comprise half the 59 available analyses and are thought the most common rock

TABLE 5.1. MATHAUNDROVE SUPER-GROUP VOLCANICS.

DATA	821	816	799	806	818	814	817	797	798	819	823	811	PJ2	810	815	822
SIO2	49.84	49.76	51.60	51.24	50.04	52.32	52.95	52.99	53.36	54.26	54.75	55.35	55.84	56.95	58.09	60.72
TIO2	0.68	0.81	0.69	1.39	0.74	0.88	0.82	0.94	1.03	1.14	1.11	1.15	0.71	0.85	0.66	0.51
AL203	13.64	15.04	14.55	15.75	15.53	17.62	19.20	17.43	17.30	15.69	15.25	16.32	18.79	16.74	18.17	18.41
FE203	9.70	9.66	9.72	10.42	11.65	10.80	9.16	9.93	9.85	12.31	11.46	11.01	3.62	8.86	7.61	5.93
FEO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	3.98	0.0	0.0	0.0
MNO	0.17	0.18	0.18	0.15	0.19	0.16	0.24	0.17	0.16	0.18	0.15	0.20	0.09	0.15	0.08	0.14
MGO	11.97	9.91	9.46	7.00	6.73	4.50	3.83	4.81	4.85	5.25	4.68	5.11	4.58	4.15	3.40	2.74
CAO	11.21	11.67	10.95	10.37	12.25	10.95	10.05	9.96	10.36	8.26	7.71	7.47	8.94	8.49	7.79	7.18
NA20	2.06	1.95	2.12	2.62	2.12	2.31	2.69	2.89	2.37	1.95	2.95	2.84	2.97	3.02	2.89	3.45
K20	0.55	0.86	0.60	0.78	0.59	0.25	0.84	0.70	0.58	0.73	1.71	0.41	0.34	0.65	1.16	0.78
P205	0.17	0.15	0.15	0.27	0.15	0.21	0.23	0.18	0.15	0.24	0.22	0.13	0.14	0.13	0.15	0.14
(LOI)	1.35	4.04	2.55	2.82	1.56	2.08	1.40	0.77	0.91	7.72	1.64	2.66	1.47	0.75	2.24	1.71
(TOTAL)	99.38	99.05	98.99	99.24	99.42	99.11	99.11	99.09	99.41	99.13	99.85	99.64	100.21	99.46	99.83	99.49
NA20/K20	3.7	2.3	3.5	3.4	3.6	9.3	3.2	4.1	4.1	2.7	1.7	6.9	8.6	4.7	2.5	4.4
MG/MG+FE2	0.75	0.71	0.70	0.62	0.58	0.50	0.50	0.54	0.54	0.51	0.50	0.53	0.58	0.53	0.52	0.53
%K	0.46	0.72	0.50	0.65	0.49	0.21	0.70	0.58	0.48	0.61	1.42	0.34	0.29	0.54	0.96	0.65
RB	16.	18.	8.7	15.	16.	9.4	12.	9.6	8.9	11.	36.	2.8	10.	6.3	26.	13.
SR	354.	377.	317.	430.	402.	760.	422.	303.	205.	1004.	258.	184.	60.	171.	361.	506.
BA	131.	295.	317.	353.	271.	299.	256.	132.	223.	423.	519.	165.	317.	81.	256.	297.
PB	1.	2.	2.	4.	2.	4.	2.	2.	2.	3.	2.	3.	0.	1.	1.	2.
K/RB	280.	400.	573.	445.	301.	219.	567.	609.	538.	556.	398.	1230.	282.	857.	375.	485.
K/SR	13.	19.	16.	15.	12.	3.	17.	19.	23.	6.	55.	19.	48.	31.	27.	13.
BA/RB	8.0	16.5	36.4	24.3	16.6	31.9	20.8	13.8	25.0	38.7	14.6	59.3	31.3	12.9	10.0	22.3
RB/SR	0.046	0.047	0.027	0.034	0.041	0.012	0.029	0.032	0.044	0.011	0.138	0.015	0.169	0.037	0.071	0.026
SR87/86	.7037	.0	.0	.7041	.7036	.0	.0	.0	.7041	.0	.0	.0	.0	.7041	.0	.0
TH	.41	.52	.31	1.2	.46	.62	.48	.52	.30	.0	.66	.31	.0	.30	.56	.31
U	.23	.19	.21	.34	.22	.32	.23	.23	.20	.0	.40	.21	.0	.23	.24	.20
TH/U	1.7	2.7	1.5	3.6	2.0	1.9	2.1	2.2	1.5	0.0	1.7	1.5	0.0	1.3	2.4	1.5
K/U	19490.	37817.	24074.	18867.	21885.	6427.	30941.	24904.	23659.	0.	35759.	16603.	0.	23099.	40785.	31545.
HPU	0.4	0.4	0.3	0.7	0.4	0.4	0.4	0.4	0.3	0.0	0.8	0.3	0.0	0.4	0.5	0.4
LA	0.0	0.0	0.0	0.0	3.1	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	16.8	0.0	0.0
YB	.0	.0	.0	.0	1.3	.0	.0	.0	.0	.0	.0	.0	.0	2.4	.0	.0
LA/YB	0.0	0.0	0.0	0.0	2.3	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	6.9	0.0	0.0
Y	19.	21.	19.	25.	19.	27.	25.	27.	32.	31.	33.	29.	0.	35.	24.	15.
NI	85.	100.	152.	116.	46.	41.	17.	29.	34.	11.	37.	12.	15.	8.	7.	9.
CO	24.	38.	39.	39.	47.	37.	25.	31.	29.	32.	33.	28.	0.	23.	22.	16.
CR	191.	332.	473.	271.	151.	71.	39.	59.	70.	9.	58.	28.	36.	10.	7.	10.
SC	20.	51.	38.	29.	47.	38.	30.	27.	38.	36.	31.	36.	0.	27.	20.	12.
V	108.	273.	262.	257.	307.	307.	265.	248.	295.	318.	288.	329.	222.	211.	183.	97.
CU	39.	82.	83.	132.	143.	159.	119.	92.	155.	294.	384.	89.	35.	36.	37.	26.
NI/CO	3.5	2.6	3.9	2.9	1.0	1.1	0.7	1.0	1.1	0.3	1.1	0.4	0.0	0.3	0.3	0.6
V/NI	1.	3.	2.	2.	7.	7.	15.	8.	9.	29.	8.	27.	15.	26.	26.	11.
MG/NI	852.	598.	374.	364.	883.	659.	1327.	984.	874.	2895.	771.	2493.	1818.	3091.	2861.	1796.
SN	.0	.0	.0	.0	.24	.0	.0	.0	.0	.0	.0	.0	.0	.25	.0	.0
HF	.0	.0	.0	.0	.65	.0	.0	.0	.0	.0	.0	.0	.0	1.4	.0	.0
ZR	43.	42.	41.	101.	39.	53.	55.	80.	75.	72.	94.	63.	0.	98.	72.	62.
B	4.	5.	0.	20.	7.	0.	0.	5.	25.	0.	11.	0.	0.	4.	0.	0.
ZR/HF	0.	0.	0.	0.	59.	0.	0.	0.	0.	0.	0.	0.	0.	69.	0.	0.
TI/ZR	96.	115.	100.	83.	114.	100.	89.	70.	82.	94.	71.	110.	0.	52.	55.	49.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

PJ2 FROM JAKES(1970)

TABLE 5.1. MATHAUNDROVE SUPER-GROUP VOLCANICS.

DATA	820	804	801	H26	B535	M96
SIO2	63.14	65.44	68.34	75.13	77.76	79.37
TIO2	0.44	0.36	0.59	0.50	0.30	0.20
AL203	16.62	17.94	15.61	13.46	11.35	10.42
FE203	6.31	4.09	4.04	2.51	1.72	2.49
FEO	0.0	0.0	0.0	0.12	0.0	0.0
MNO	0.19	0.09	0.05	0.01	0.03	0.05
MGO	2.37	1.52	0.78	0.34	0.81	0.28
CAO	6.69	5.72	5.62	2.45	2.50	1.76
NA20	3.21	3.84	3.38	4.71	4.01	4.08
K20	0.76	0.85	1.47	0.64	1.45	1.17
P205	0.28	0.15	0.12	0.14	0.07	0.18
(LOI)	1.64	2.62	1.08	1.85	1.66	2.32
(TOTAL)	99.47	98.98	98.96	99.95	99.56	100.06
NA20/K20	4.2	4.5	2.3	7.3	2.8	3.5
MG/MG+FE2	0.48	0.47	0.32	0.24	0.53	0.21
%K	0.63	0.71	1.22	0.53	1.20	0.97
RB	6.2	17.	23.	8.2	12.	6.2
SR	442.	556.	224.	108.	94.	97.
BA	235.	322.	245.	90.	138.	123.
PB	2.	3.	2.	1.	4.	3.
K/RB	1007.	425.	520.	654.	982.	1551.
K/SR	14.	13.	55.	49.	128.	100.
BA/RB	37.7	19.4	10.4	11.0	11.2	19.7
RB/SR	0.014	0.030	0.105	0.075	0.130	0.064
SR87/86	.7041	.0	.0	.0	.7043	.7042
TH	.52	1.0	1.1	.45	.0	.41
U	.25	.26	.63	.41	.0	.57
TH/U	2.1	4.0	1.8	1.1	0.0	0.7
K/U	25596.	27228.	19280.	13075.	0.	16899.
HPU	0.5	0.6	1.0	0.5	0.0	0.8
LA	5.2	0.0	0.0	60.9	0.0	6.8
YB	1.8	.0	.0	28.	.0	2.9
LA/YB	2.8	0.0	0.0	2.2	0.0	2.4
Y	26.	117.	33.	560.	34.	52.
NI	3.	7.	4.	2.	ND	ND
CO	11.	8.	8.	0.	ND	ND
CR	4.	10.	12.	0.	3.	3.
SC	8.	7.	15.	0.	7.	9.
V	82.	74.	127.	0.	7.	0.
CU	11.	19.	13.	0.	7.	6.
NI/CO	0.3	0.9	0.5	0.0	0.0	0.0
V/NI	27.	10.	31.	0.	0.	0.
MG/NI	4665.	1258.	1146.	995.	0.	0.
SN	.21	.0	.0	.35	.0	.28
HF	1.2	.0	.0	2.2	.0	2.0
ZR	84.	71.	134.	177.	142.	119.
B	0.	0.	5.	0.	0.	0.
ZR/HF	68.	0.	0.	79.	0.	58.
TI/ZR	31.	31.	27.	17.	13.	10.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

H26 MAJOR AND SOME TRACE ELEMENT DATA FROM HINDLE(1970)

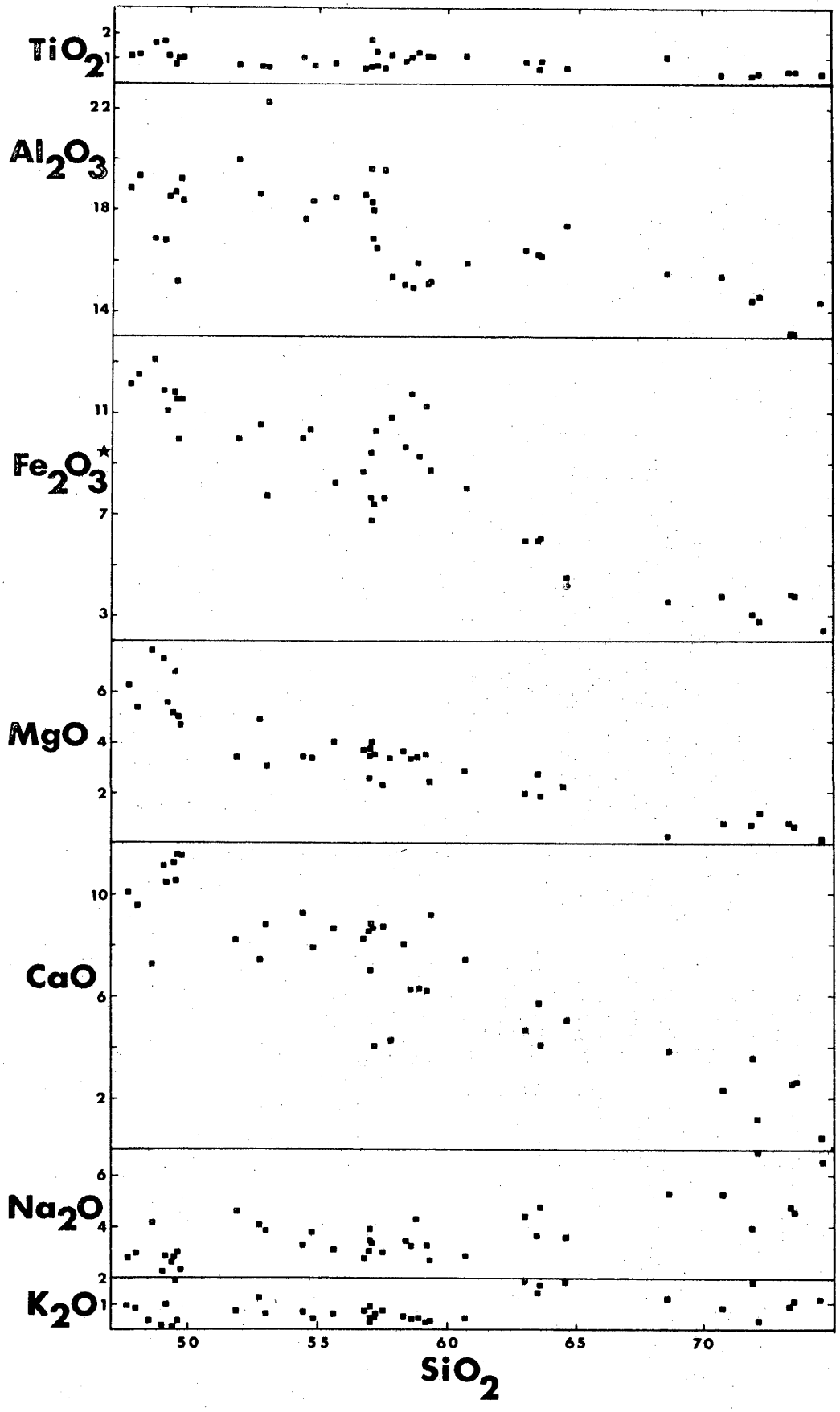

Figure 5.2 Major element oxides of the Mathuandrove Super-Group, Vanua Levu. Data from Table 5.1, Rodda (1969), and Hindle (1970 and pers. comm., 1971).

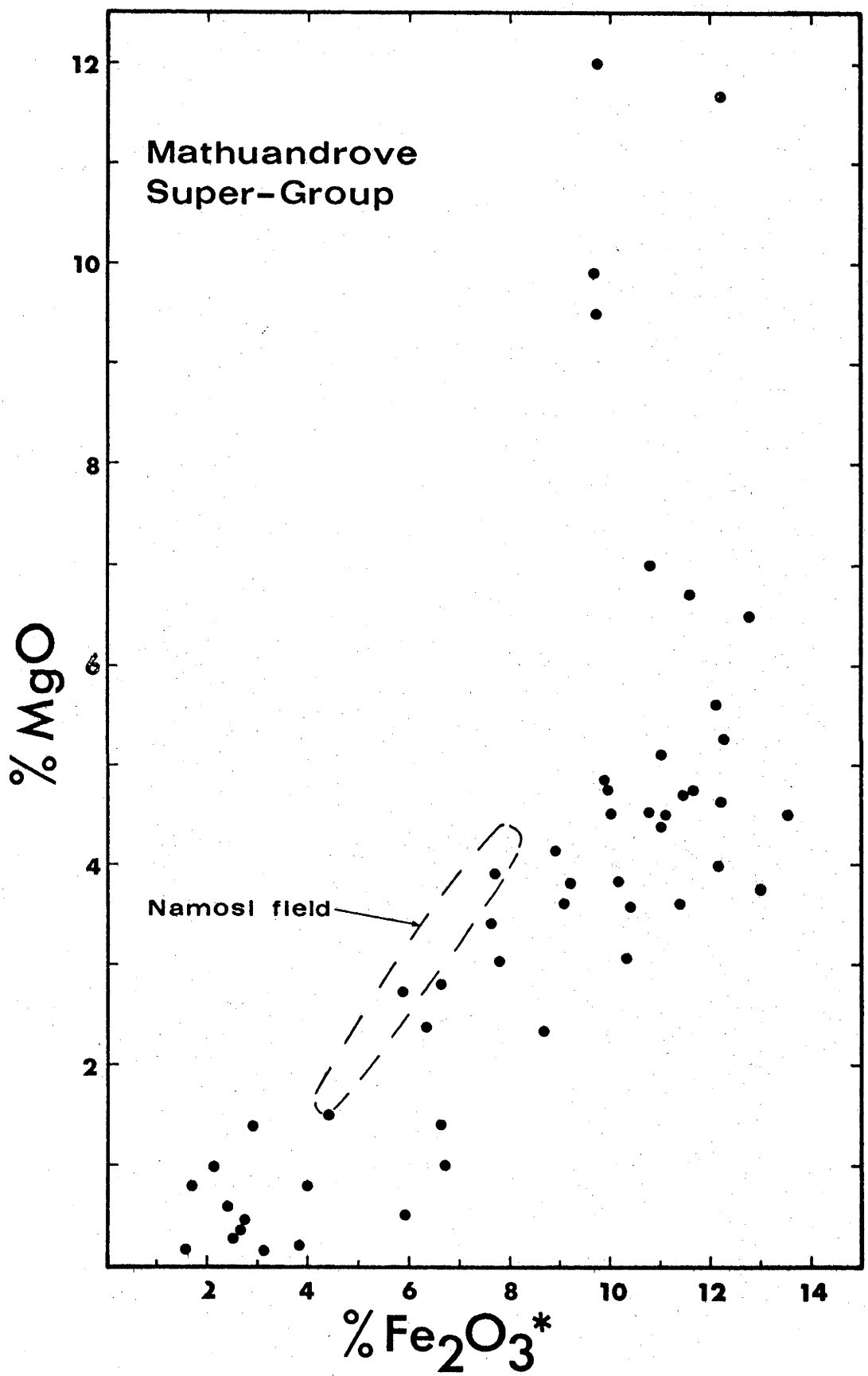

Figure 5.3 Fe-Mg variations in Mathuandrove Super-Group samples, Vanua Levu. Data sources as in Figure 5.2. Namosi Andesite field reproduced from Figure 4.3.

type in outcrop, have an average of 0.7% K_2O, 11 ppm Rb, 370 ppm Sr with 87/86 ratios of 0.7041, 230 ppm Ba, 2 ppm Pb, 0.5 ppm Th, and 0.23 ppm U. Iron enrichment is extremely variable (9-13.5% Fe_2O_3* between 51-55% SiO_2 or 8-13.5% Fe_2O_3* at 4.5% MgO) and without relationship to alkali contents. TiO_2 contents increase from 0.85 to 1.2% as SiO_2 increases to 56%, but decrease gradually in more siliceous rocks. The following are average ratios for these basaltic-andesites and andesites: Na_2O/K_2O = 4.7; K/Rb = 625; Ba/Rb = 26; Rb/Sr = 0.038; Th/U = 2.0; and K/U = 25,000. Co, Sc, and V contents decrease by a factor of three with increasing SiO_2 and Ni and Cr by factors of five and seven, respectively, from about 50 ppm Ni, 40 ppm Co, 70 ppm Cr, 40 ppm Sc, and 300 ppm V, while Ni/Co ratios remain ≲1, V/Ni increase from 7 to 27, and Mg/Ni from 600 to 3000. One REE pattern is presented in Figure 5.4 for andesite 810. It is moderately enriched in light REE (La/Yb = 7), has a slightly negative Eu anomaly (Eu/Eu* = 0.79), and heavy REE abundances at eleven to twelve times their chondritic levels. The intrusive Naroro Hornblende Andesite plugs, younger than most analyzed volcanics, are indistinguishable in composition from other MSG acid andesites.

Table 5.2 Rare earth analyses of Mathuandrove Super-Group samples, Vanua Levu

	818	810	820	H26	M96
La	3.0	7.0	5.1	60.	6.6
Ce	7.4	16.6	10.7	103.	16.8
Pr	1.4	2.4	2.5	41.	3.4
Nd	7.2	11.2	11.5	230.	17.2
Sm	1.7	2.7	2.7	67.	6.2
Eu	.66	.78	.89	7.7	1.7
Gd	2.0	3.7	-	-	5.2
Tb	.37	.54	.45	11.6	.90
Dy	2.2	3.4	2.8	6.1	5.2
Ho	.54	.92	.66	15.3	1.3
Er	1.4	2.4	1.8	36.	3.4
Yb	1.3	2.4	1.8	27.	2.8

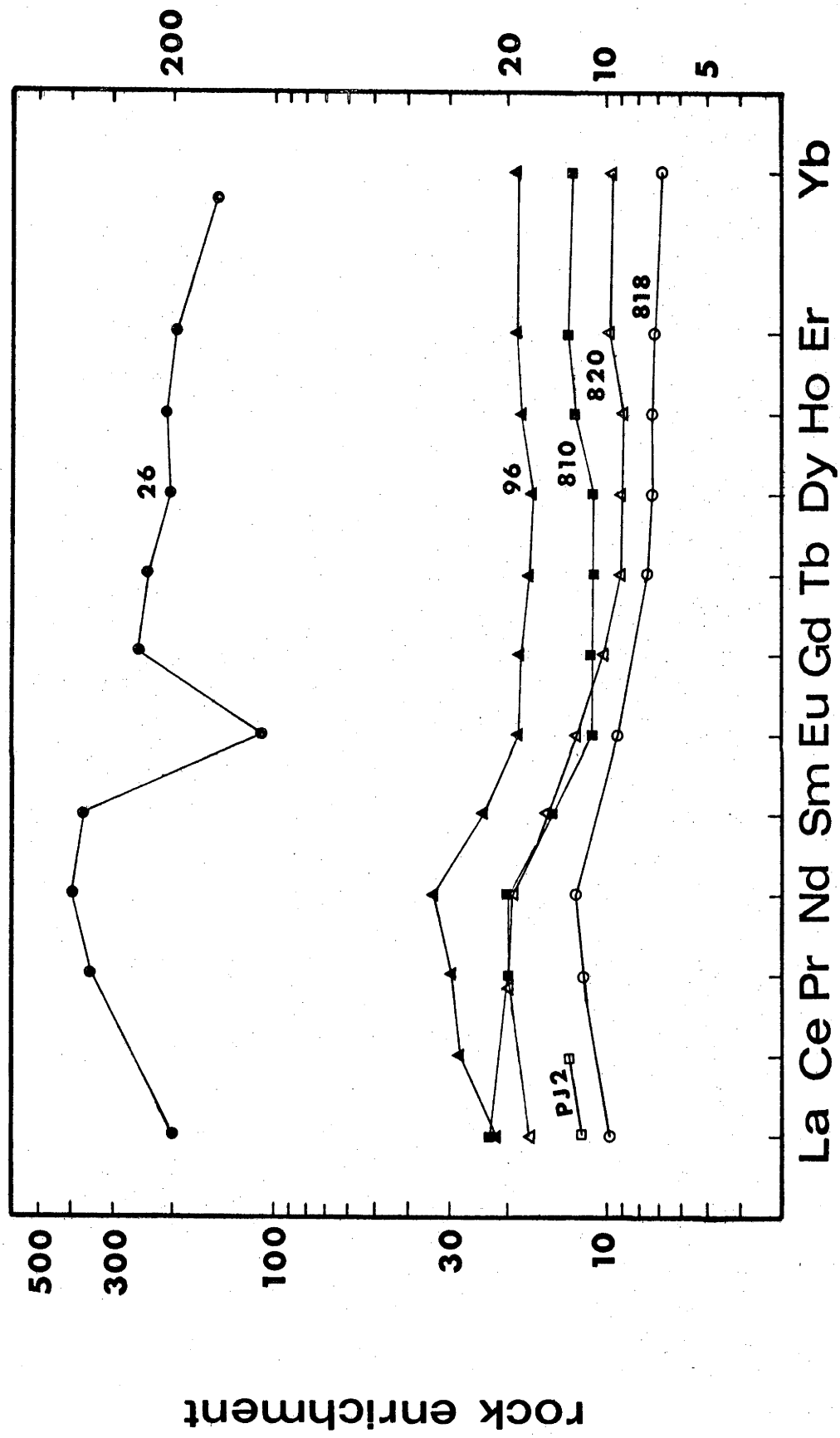

Figure 5.4 Rare earth patterns of Mathuandrove Super-Group samples, Vanua Levu. Data from Table 5.2 and Jakes (1970).

Dacites 904 in Table 5.1, 19 from Hindle (1970), and AN6 from Rodda (1969 are more silicic variations on this theme with comparable or slightly greater concentrations of K, Rb, Sr, Ba, Pb, U, and Th, and similar large cation ratios.

The rhyolites from Udu Peninsula also share many of these compositional features. The five analyzed samples have >75% SiO_2 yet an average of only 1.0% K_2O, 9 ppm Rb, 100 ppm Sr with 87/86 ratios of 0.7042, 120 ppm Ba, 3 ppm Pb, 0.4 ppm Th, and 0.5 ppm U. They have an average K/Rb ratio of 1060, Ba/Rb of 14, K/Sr of 92, and Rb/Sr of 0.09. Sample 96 has a REE pattern enriched in Pr, Nd, and Sm (La/Sm = 1.05), no Eu anomaly, and both light and heavy REE enriched about twenty times relative to chondrites (La/Yb = 2.4). These REE features are much more accentuated in sample 26 (major element data from Hindle, 1970) in which Pr, Nd, and Sm are 300 to 400 times enriched relative to chondrites, La/Sm and La/Yb ratios are 0.9 and 2.2 respectively, and La and heavy REE are enriched about 200 times relative to chondrites. It also has negative Ce and Eu anomalies (Eu/Eu* = 0.34), and is one of the few terrestrial analogues to the REE patterns of lunar KREEP.

The relationship of analyzed basalts to the intermediate and acidic rocks is unclear. Samples 816, 821, 818, 806, and 821 are tholeiitic basalts about critically saturated with quartz if $Fe_2O_3/FeO = 0.6$ (821 and 816 have six and three per cent ol, respectively; 799, 806, and 818 have two to three per cent qz). Nevertheless they are from four units sufficiently different to be mapped as separate formations: 806 and 799 are undifferentiated Natewa Group rocks from central Vanua Levu; 818 is from the Wairuku Formation type locality (see Woodrow in FGSD Annual Report for 1968, p. 5); 821 is from the older Savundrondro Formation (ibid.); and 821 is from the Dakuniba area of Cakaudrove Peninsula. (Rocks from this latter site were once thought younger than those I refer to as constituting the Mathaundrove Super-Group [Rickard, 1966] and are now considered stratigraphically ambiguous [Woodrow, loc. cit.]. Sample 821 clearly belongs to the MSG, not Mbua Group, from which it differs in normative mineralogy, 87Sr/86Sr ratios, and such diagnostic minor and trace elements as Ti, Zr, P, Th, and U.) These five basalts are similar to most MSG basaltic-andesites and andesites in incompatible minor and trace element contents other than Rb which, except in the case of sample 799, is almost twice as abundant and results in lower K/Rb and higher Rb/Sr ratios in the basalts. In principal, therefore, these

basalts could yield the intermediate rocks observed through crystal fractionation, although some large cation contents--notably K, Rb, Ba, and Th--seem variable.

Some exceptions to this geochemical pattern, especially in concentrations of the large cations, seem geographically consistent. These elements appear to have been enriched by some process operating principally in central Vanua Levu (see samples 801 and 823 in Table 5.1 and samples 12, 13, 20, 23, 24, and 25 in Hindle [1970]). Hindle's analyses of rocks from Kia, a small islet 22 km north of central Vanua Levu, are similarly alkali-rich at 67% SiO_2. K, Rb, Ba, Th, and U are uniformly enriched; Sr and most ferromagnesian trace element contents are no different from those of other Mathaundrove Group samples. Neither are other major elements. Cu and Zr seem to have been concentrated with the large cations although Cu remains negatively correlated with Si as in the larger suite. If these rocks are spatially constrained they could define a unique eruption, or local but intermittent volcanic event. More control is necessary if resolution is sought.

Another possible area of local variability is near Savusavu. Basalts 818 and 821 and dacite 820 have 87Sr/86Sr ratios lower than those observed elsewhere in Vanua Levu. Complete REE patterns for samples 818 and 820 and a partial one for sample 2 from Jakes (1970) are given in Figure 5.4. All are generally flat relative to chondrites (La/Yb ratios of 2.3 for the basalt and 2.8 for the dacite) with slight enrichments of Pr and Nd, and slight (10 to 15%) positive Eu anomalies scarcely exceeding analytical uncertainties. This Pr and Nd enrichment differs from the REE pattern of andesite 810 from northern Vanua Levu but presages those of Udu Peninsula rhyolites. Dacite 820 also has lower heavy REE contents than andesite 810 despite being more fractionated from a basaltic composition in most other elements.

D. Summary and comparison to the older Viti Levu island arc tholeiitic series

The Mathaundrove Super-Group rocks record largely submarine Late Miocene to Early Pliocene volcanism which yielded mostly though not uniformly an island arc tholeiitic series. It shares the following compositional features with the tholeiitic series erupted during the first period of volcanism on Viti Levu: wide SiO_2 range; TiO_2 about one per cent in basalts and basaltic-andesites; moderate iron-enrichment;

low K, Rb, Ba, Sr, Pb, Th, and U contents throughout the suite; REE patterns generally flat or slightly inflected relative to chondrites (La/Yb <3); and a predominance of basaltic-andesite. In detail these features differ, rendering the appellation "island arc tholeiitic series" less appropriate to the Vanua Levu suite as large cations are more abundant, K/Rb ratios are lower, and Rb/Sr and La/Yb ratios generally higher there. Vanua Levu basalts have higher maximum MgO and ferromagnesian trace element contents as well as $Mg/Mg+Fe^{+2}$ ratios. As a suite, the Vanua Levu rocks contain more Ni and Cr at similar SiO_2 or MgO contents and Mg/Ni ratios are lower.

There are exceptions to both suites, principally in K, Rb, Ba, Th or U concentrations. These are better documented and appear to have some spatial consistency in Vanua Levu.

Magmas in both areas precipitated plagioclase, clinopyroxene, and minor Fe-Ti oxide. Amphibole is uncommon except in late intrusive plugs or dikes where P_{H2O} could increase locally. Hypersthene is common in Vanua Levu, rare in Viti Levu. Pigeonite is not recognized in the groundmass of either suite.

Magmas evolved differently in various areas of Vanua Levu during the Late Miocene to Early Pliocene but most rocks appear to share a common genetic heritage despite their geographical distribution or stratigraphic position within the MSG.

5.3 Mbua Basaltic Group

A dissected olivine-basalt shield volcano in southwestern Vanua Levu was inferred seventy years ago by Guppy and shown to be stratigraphically younger than rocks of the Mathuandrove Super-Group by Rickard (1966), Coulson (1971), and Hindle (in prep.). McDougall and Hindle, in their project mentioned above, should determine its radiometric age; on geomorphic grounds it is unlikely to be older than a few million years.

Petrographic and minerographic details are available from Ibbotson (in Coulson, 1971, Table 2). Labradorite, augite, olivine, and an oxide are common phenocrysts and usually in that order of abundance although local variants occur.

Three new analyses are given in Table 5.3. Rodda (1969) compiled two, and Hindle (1970) presents several others including some of silicic (>60% SiO_2) differentiates. They are tholeiitic basalts (see Figure 1.3)

TABLE 5.3. MBUA VOLCANIC GROUP.

DATA	807	795	808
SIO2	48.67	49.10	49.07
TIO2	2.23	2.25	2.26
AL203	18.28	16.78	16.91
FE203	10.41	11.04	10.93
FEO	0.0	0.0	0.0
MNO	0.17	0.15	0.15
MGO	5.19	6.03	5.82
CAO	10.60	10.41	10.84
NA20	3.08	3.05	2.97
K20	0.96	0.81	0.67
P205	0.41	0.37	0.37
(LOI)	2.00	1.59	1.36
(TOTAL)	99.56	98.94	99.87
NA20/K20	3.2	3.8	4.4
MG/MG+FE2	0.55	0.57	0.56
%K	0.80	0.67	0.56
RB	13.	14.	8.1
SR	535.	466.	500.
BA	226.	205.	193.
PB	2.	3.	3.
K/RB	600.	468.	685.
K/SR	15.	14.	11.
BA/RB	16.9	14.3	23.7
RB/SR	0.025	0.031	0.016
SR87/86	.0	.0	.7041
TH	2.3	1.7	1.9
U	.44	.41	.44
TH/U	5.1	4.2	4.4
K/U	18147.	16395.	12742.
HPU	1.0	0.8	0.9
LA	0.0	0.0	0.0
YB	.0	.0	.0
LA/YB	0.0	0.0	0.0
Y	29.	29.	28.
NI	50.	78.	73.
CO	40.	37.	39.
CR	64.	150.	154.
SC	19.	25.	23.
V	194.	205.	215.
CU	37.	53.	49.
NI/CO	1.3	2.1	1.9
V/NI	4.	3.	3.
MG/NI	623.	466.	480.
SN	.0	.0	.0
HF	.0	.0	.0
ZR	164.	156.	153.
B	0.	0.	0.
ZR/HF	0.	0.	0.
TI/ZR	82.	86.	89.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

with several percent normative hyp. Hindle's differentiates are strongly qtz-normative. I have included sample 795 as Mbua Basalt because of its composition and youthfulness even though it is not spatially contiguous with the others. It intrudes and is therefore younger than Mathuandrove Super-Group rocks east of Labasa. Other young olivine basalt intrusives are known elsewhere in central Vanua Levu (Ibbotson, 1969, pp. 19, 23, and 31).

These differ strongly from most basalts in island arc environments (see Chayes, 1964b), including those of third period Viti Levu volcanism, in their relatively high Ti (>2%), Zr (>130 ppm), and Th (>2 ppm) contents. The one 87Sr/86Sr ratio measured was 0.7041.

Thus Vanua Levu volcanism changed considerably with time, from predominantly intermediate to basaltic. The basalts themselves are also different: both are tholeiitic but those of the Mbua Basaltic Group contain more calcic normative plagioclase, less hyp and di, have twice as much Ti and Zr, more Sr, Th, U, and 87Sr (but not Rb), and less Cu than basalts of the Mathuandrove Super-Group. (Sample 806, collected from an outcrop of presumed MSG rocks, seems transitional between the groupings.) The Mbua intermediate rocks analyzed by Hindle are extremely different from MSG samples of comparable silica contents. The mean Cu content of Mbua Basalts is 46 ppm and of MSG basalts to andesites is 120 ppm. Coulson (1971) reports that "background" Cu concentrations are 40 and 150 ppm in stream sediment samples taken from Mbua and Mathuandrove Super-Group outcrop areas, respectively.

Chapter 6. Geology and Geochemistry of the Lau Islands

6.1 Introduction

The islands and atolls of Lau are small pinnacles of a submarine ridge to which they give their name. This ridge extends from 16°S to the Coromandel Peninsula of New Zealand's North Island, rises 4 km above the adjacent South Fiji Basin, averages about 150 km in width at depths of 2 km, and appears to be a continuous bathymetric feature rather than a mélange of disparate crustal blocks (see Figure 1.1 and Chase et al., 1968). Viti Levu and Vanua Levu are simply its most and second-most emergent portions.

This Lau-Colville Ridge has attracted attention recently as a boundary of one of the most studied marginal seas (Karig, 1970) and as the only emergent "third arc" in the world. One crustal profile by Shor et al. (1971) at 31°S suggests that 15 km of crust, half of it a "thickened oceanic layer" with 7.0 km/sec velocities, underlies the ridge. Apparently, therefore, the ridge thins southward. Mantle beneath it has low seismic attenuation (Q ∿1000: Barazangi and Isacks, 1972) with V_p = 8.0 km/sec (Shor et al., 1971), in sharp contrast to mantle beneath the adjacent Lau Basin.

The Lau islands have not yet been geologically mapped by the FGSD but were visited by Dana, Agassiz, and others as early as the nineteenth century and figured somewhat in the controversy over the origin of coral reefs and atolls. The most comprehensive geological study was conducted by Ladd and Hoffmeister in 1934 which resulted in several papers concerning geology, petrography, limestone composition, and paleontology (Ladd and Hoffmeister, 1945; Kleinpell, 1954; and Johnson and Ferris, 1950). Their summary of volcanism is similar to that given in chapter 5 for Vanua Levu: initially submarine Neogene volcanism of intermediate composition (the Lau Volcanics) followed by younger olivine basalts (the Korombasanga Volcanics).

This history is best exposed in the Exploring Islands, named after Wilkes' U.S. Exploring Expedition which, in 1840, charted the islands and brought to them J.D. Dana, of geosynclinal theory fame, as ship's geologist. Most of my samples and observations were collected and made in these islands although I also collected samples from Katafaga and Cicia and was given one from Ono-i-Lau. Locations of the 27 Lauan rocks analyzed are indicated in Figures 6.1 and 6.2 and listed together with petrographic descriptions in Appendix 2. Most of the geological and

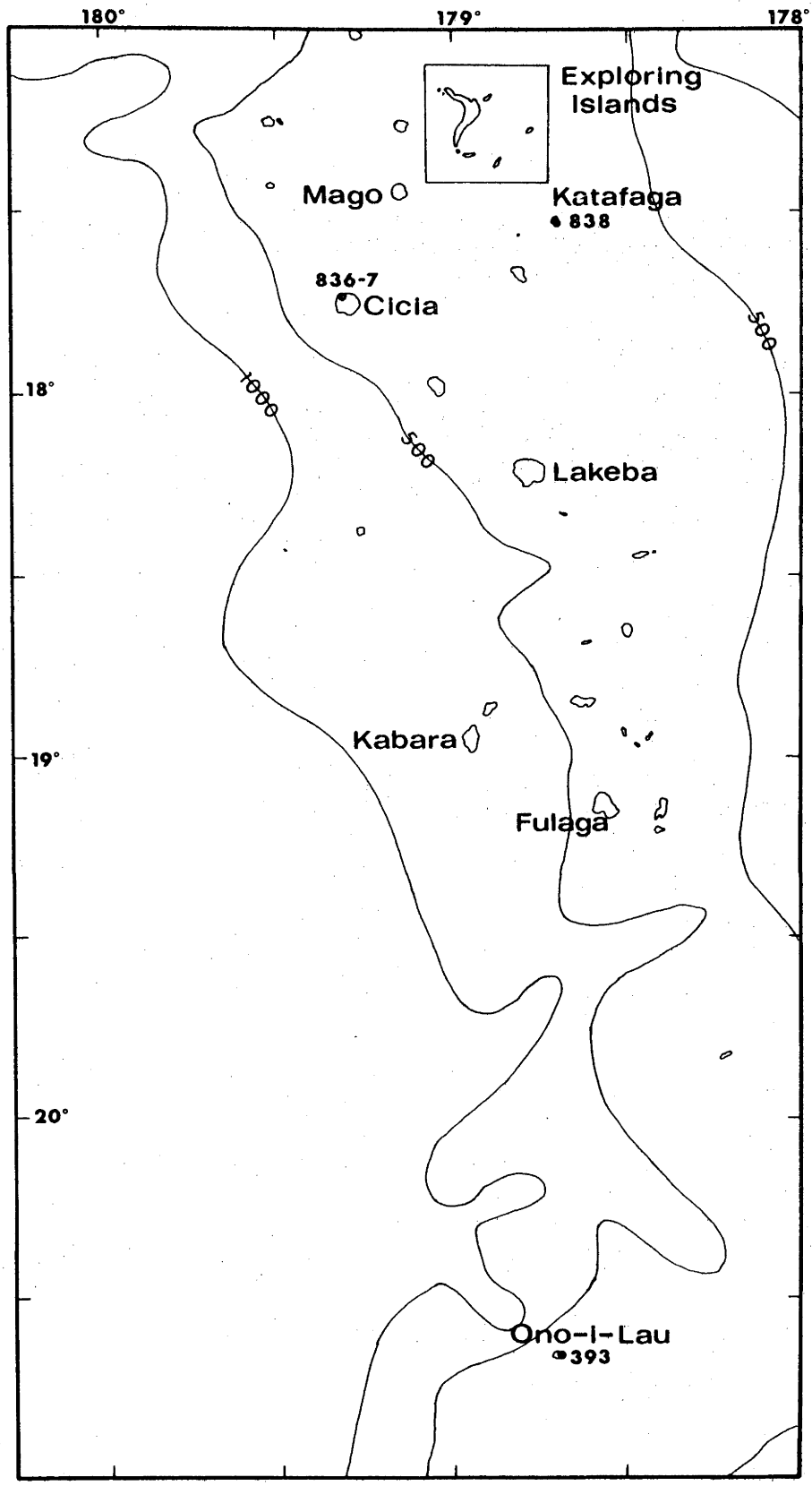

Figure 6.1 The Lau Islands. Bathymetry from Phillips (1964). Inset refers to area covered by Figure 6.2.

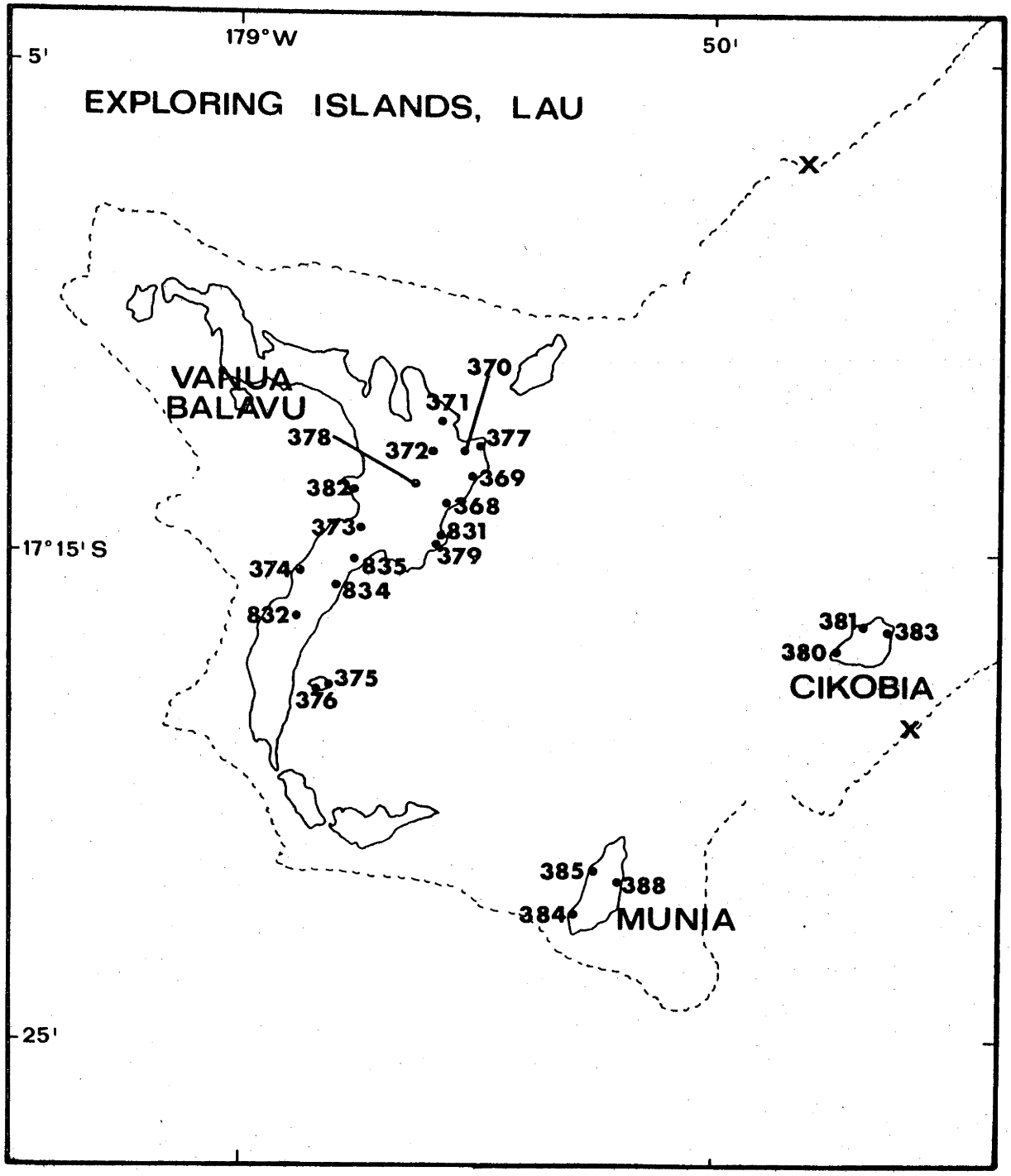

Figure 6.2 Location of samples from the Exploring Islands, Lau. Dashed lines indicate the fringing reef position; x indicates good fishing spots.

some of the petrographic descriptions which follow are from Ladd and Hoffmeister, supplemented or altered by my observations.

6.2 The Lau Volcanics: Late Miocene

A. Environment, age, and distribution

The Lau Volcanics are predominantly agglomerates and tuffs with some flows a few of which developed lava pillows. The tuffs are often bedded and indicate shallow marine deposition.

Ladd and Hoffmeister (1945) and Kleinpell (1954) consider the Lau Volcanics Lower Miocene or older (i.e. ≧12.5 m.y.: Figure 1.4) because they are overlain by the Futuna Limestone which contains a larger foram and mollusk fauna of lower Tertiary f age. However, data presented in Table 6.1 indicate the formation is, instead, only 7.5 to 9.0 m.y. old. This discrepancy is important both for establishing the correlation of Lau volcanism with that elsewhere in Fiji and in Tonga, and for dating the maximum age of the Lau Basin (see chapter 8).

Table 6.1 K-Ar ages for Lau samples

No.	Sample	K%	Rad.Ar^{40} ($x10^{-6}$cc NTP/g)	Rad.Ar^{40} (%)	Calculated Age (m.y.)
		LAU VOLCANICS			
834	Whole rock	0.900	0.323	62.7	9.0 ±0.13
	Vanua Balavu	0.886			
835	Whole rock	1.265	0.419	37.9	9.3 ±0.17
	Vanua Balavu	1.264			
837	Whole rock	0.964	0.288	46.9	7.5 ±0.17
	Cicia	0.971			
838	Whole rock	0.518	0.162	28.0	7.8 ±0.19
	Katafaga	0.516			
		KOROMBASANGA VOLCANICS			
831	Whole rock	0.520	0.0752	19.2	3.5 ±0.17
	Vanua Balavu	0.545			
393	Whole rock	3.172	0.744	59.4	5.9 ±0.09
	Ono-i-Lau	3.157			

Four Lau Volcanic whole rock samples, two from Vanua Balavu and one each from Katafaga and Cicia, were dated with the help of I. McDougall and Z. Roksandic in a brief reconnaissance study. (I had hoped to extend it for this thesis but was prevented by lack of time.)

All have a very fine-grained groundmass with quench crystals of plagioclase, pyroxene, and an opaque set in a glassy matrix. Only in sample 837 did this glass appear devitrified. Potassium determination by flame photometry were not of high quality and consistently exceeded by about 6% those determined for different aliquots of the same sample by XRF and gamma-ray spectrometry. This could result in underestimates of age by 0.5 m.y. Ar analyses were good with only 40 to 70% atmospheric Ar present and isotope ratio measurements having coefficients of variation of 0.1 to 0.5%.

The Futuna Limestone contains larger foraminifera diagnostic of the lower f (f_{1-2}) stage (compare the list from W.S. Cole in Ladd and Hoffmeister, 1945, Table 18, with the range charts of Adams, 1970 and Clarke and Blow, 1969), equivalent to planktonic zones N.9 to 12 (see Table 1.4). It also contains planktonic forms (i.e. Globorotalia [G.] multicamerata [=G. menardii var fijiensis], Globigerina conglomerata and G. bulloides: Kleinpell, 1954) indicative of zones N.18 or 19 (Blow, 1969). The benthonic forms are most common in lower tuffaceous horizons above reworked Lau Volcanics and are much less common than the younger planktonic varieties in stratigraphically higher sections where the formation is, in fact, described as a "Globigerina limestone" (Crickmay et al., 1941). (Veizer, 1972, analyzed another Futuna Limestone sample which was almost pure $CaCO_3$ with an 87Sr/86Sr ratio of 0.7093, confirming Crickmay et al.'s argument for primary, shallow [~80m] deposition).

The same paleontological inconsistency is found in fauna of the Suva Marl with which at least the lower Futuna Limestone has been correlated (Ladd and Hoffmeister, 1945; Kleinpell, 1954). It contains a similar f_{1-2} larger foram fauna but also planktonic forms indicative of zones N.17 to 19 (Blow, 1969, Figure 37). It is even a paratype locality for zone N.18 which straddles the Miocene-Pliocene boundary (5.5 m.y.: Berggren, 1969). Tholo plutonic detritus is found within formations laterally equivalent to the Suva Marl (Rodda, 1967). If the age of this plutonism or its uplift is 7 to 13 m.y. (see section 4.5A), an f_{1-2} age for the marl is impossible on the time scales of Berggren or Page and McDougall. (If 7 to 13 m.y. is only the time of Namosi volcanism when Tholo pluton K-Ar systems were reclosed, this argument could still but will not necessarily hold.)

Planktonic and plutonic evidence, therefore, strongly suggests the Suva Marl is Tertiary g in age. This requires that its larger foraminifera, largely on the basis of which the Suva Marl and Futuna Limestone were

originally correlated, are either more diachronous than presently thought or are reworked. Rodda (written comm., 1970) believes the latter and notes some overlap between Suva Marl and Wainimala Group faunas. Ladd (written comm., 1971) regards the larger forams of the Suva Marl too plentiful and well-preserved to be reworked. Cole (in Adams, 1970, p. 124) apparently applies the same reasoning to the Futuna Limestone assemblage for he cites one of the plankton-bearing horizons mentioned above as an instance of Orbulina occurring with an f_{1-2} fauna which, Adams notes, is rare.

My K-Ar measurements suggest the Suva Marl and Futuna Limestone are correlative after all, but Tertiary g instead of f_{1-2} in age. This appears consistent with the planktonic foram fauna of both formations If so, this means Lau volcanism was contemporaneous with that which produced the Namosi Andesites of Viti Levu and the Mathuandrove Super-Group rocks of Vanua Levu, and that formations similar in age to those of the Viti Levu basement lie unexposed beneath the Lau Volcanics. I shall adopt this view hereafter.

Alternatively, my K-Ar results may be spurious, the planktonic forams anomalous, and the Suva Marl-Futuna Limestone correlation mistaken. If so, Lau volcanism may have been contemporaneous with the last of Viti Levu's first period activity or, indeed, altogether unrelated. Additional radiometric dating is planned to resolve the issue.

B. Petrography and stratigraphic identification

The Lau Volcanics are basaltic andesites, andesites, and dacites which precipitated plagioclase, two pyroxenes, and minor amounts of an oxide upon cooling. Pyroxene phenocrysts usually comprise 0 to 15% of the mode and are either augite, hypersthene, or both; their relative proportion is variable and not a function of bulk composition in my samples. Plagioclase, often ranging from bytownite to andesine within single phenocrysts, is usually more abundant than pyroxene. Hornblende is absent except for one hornblende-gabbro xenolith in sample 377 (hornblende crystal is 1.2 x 0.5 mm where hornblende has reacted with liquid to give clinopyroxene and an opaque in a thick [0.2mm] reaction rim). The groundmass is usually very fine-grained and glassy containing quench crystals of plagioclase, pyroxene, and an opaque. Alling (in Ladd and Hoffmeister, 1945, p. 196) describes the groundmass pyroxene as pigeonite ($2V < 40°$) but I was unable to confirm this.

Most samples included as Lau Volcanics in Table 6.2 were collected from either outcrops or outcrop areas mapped as such by Ladd and Hoffmeister. Others were so designated for petrographic reasons even when they had cropped out in areas mapped as containing Korombasanga Volcanics by those authors. These include samples 373 and 379 from Vanua Balavu, 383 from Cikobia-i-Lau, and all from Munia. Subsequent analyses vindicated these decisions. I found no rocks likely to be Korombasanga Volcanics on these latter two islands although they were inferred to occur on Cikobia-i-Lau for essentially geomorphic reasons (Ladd and Hoffmeister, 1945, pp. 48-9) and on Munia because the one sample examined by Alling (op. cit., pp. 206-7) and the five collected by Foye (1918, pp. 125-6) contained olivine phenocrysts.

C. Geochemistry

Nineteen new analyses of Lau Volcanics are given in Tables 6.2 and 6.3 and depicted in Figures 6.3 to 6.5. Foye (1918) gives two others. These define a suite with moderate SiO_2 range, ∿1% TiO_2, moderate iron enrichment, generally high Al_2O_3 (17.6 ±1.4%), about 1% K_2O, and Na_2O/K_2O ratios of 3.2 ±0.7. Most $Mg/Mg+Fe^{+2}$ ratios are within the range 0.55 to 0.45; average ratios for andesites are similar to those for basaltic-andesites.

Although similar in major element components, two statistically and spatially distinct groups of analyses are conjoined in Table 6.2. One consists of samples 380, 388, 383, 385, 838, 381, and 384 which are from the islands of Cikobia-i-Lau, Munia, and Katafaga on the eastern edge of the Lau-Colville Ridge. The others are from Vanua Balavu and Cicia to the west. (This is a crude geographic distinction as Katafaga is further east of Munia and Cikobia-i-Lau, and Cicia further west of Vanua Balavu, than these Exploring Islands are distant from one another; see Figure 6.1.) The eastern group has an average of 10 ppm Rb, 310 ppm Sr, 165 ppm Ba, 0.5 ppm Th, and 0.3 ppm U whereas the western group has an average of 21 ppm Rb, 360 ppm Sr, 210 ppm Ba, 0.8 ppm Th, and 0.5 ppm U. Average K/Rb ratios for the eastern and western groups, respectively, are 740 and 510; Ba/Rb ratios, 16 and 10; Rb/Sr ratios, 0.037 and 0.063; and Th/U ratios, 1.4 and 2.0.

Two REE analyses for each group are given in Table 6.3 and Figure 6.5. Those from the eastern group have considerably higher REE contents, i.e. are more enriched in all but especially heavy REE relative to chondritic abundances. All four samples have La/Yb ratios

TABLE 6.2. LAU VOLCANICS.

DATA	379	380*	374	837	388*	369	383*	385*	370	834	838*	832	836	377	835	381*	373	384*	375
SIO2	54.95	55.00	55.34	55.49	55.87	55.94	56.59	56.64	56.96	57.10	57.16	57.21	58.08	59.03	59.63	60.38	60.92	64.35	66.24
TIO2	0.63	0.91	0.96	0.88	0.97	0.73	0.90	0.93	0.70	0.95	0.92	0.92	0.91	0.59	0.99	0.80	0.90	0.88	0.97
AL203	20.75	17.48	17.33	15.44	19.05	19.25	17.84	16.86	19.43	17.50	16.95	17.19	18.54	17.69	17.56	16.98	16.90	14.84	15.85
FE203	7.08	9.72	8.90	8.38	8.28	7.77	8.71	8.75	7.23	8.35	8.99	8.56	7.68	7.22	6.86	7.28	7.09	7.02	5.72
FEO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
MNO	0.19	0.19	0.17	0.15	0.15	0.21	0.18	0.16	0.20	0.17	0.16	0.17	0.10	0.24	0.17	0.18	0.14	0.16	0.08
MGO	2.87	4.15	4.63	6.50	2.53	3.08	3.29	4.24	2.51	3.58	3.77	3.70	2.03	3.08	2.43	2.78	2.40	1.96	0.51
CAO	9.20	8.67	8.63	8.56	8.82	8.44	8.33	8.16	8.11	7.77	8.29	7.89	6.61	7.15	7.10	6.81	6.35	5.39	4.02
NA20	2.93	2.87	2.83	3.24	3.12	3.02	3.16	3.27	3.39	3.38	3.02	3.15	4.45	3.44	3.51	3.64	3.09	3.61	4.18
K20	1.08	0.84	1.00	1.09	0.98	1.20	0.81	0.79	1.07	0.98	0.61	1.04	1.29	1.10	1.45	0.96	1.95	1.52	2.12
P205	0.32	0.16	0.21	0.27	0.23	0.35	0.18	0.20	0.39	0.22	0.12	0.16	0.29	0.46	0.28	0.19	0.26	0.26	0.32
(LOI)	1.41	0.77	0.81	1.10	0.82	0.65	0.93	0.66	0.85	0.44	1.22	1.19	1.73	1.27	1.52	1.73	2.33	0.65	3.71
(TOTAL)	99.61	99.25	99.56	99.18	99.01	99.45	99.53	99.23	99.09	99.29	99.29	99.30	100.63	99.03	100.02	99.93	99.35	100.35	98.70
NA20/K20	2.7	3.4	2.8	3.0	3.2	2.5	3.9	4.1	3.2	3.4	4.9	3.0	3.4	3.1	2.4	3.8	1.6	2.4	2.0
MG/MG+FE2	0.50	0.51	0.56	0.65	0.43	0.49	0.48	0.54	0.46	0.51	0.50	0.51	0.39	0.51	0.46	0.48	0.45	0.40	0.18
%K	0.90	0.70	0.83	0.91	0.81	1.00	0.67	0.66	0.89	0.81	0.51	0.86	1.07	0.92	1.21	0.79	1.62	1.27	1.76
RB	19.	7.8	15.	27.	12.	19.	8.4	8.9	20.	15.	10.	16.	21.	20.	22.	9.0	26.	18.	28.
SR	509.	508.	269.	367.	321.	492.	291.	308.	506.	284.	254.	244.	349.	479.	285.	282.	272.	245.	245.
BA	275.	102.	116.	282.	132.	202.	122.	127.	204.	152.	345.	182.	239.	225.	213.	153.	165.	201.	242.
PB	2.	2.	1.	2.	1.	2.	2.	1.	2.	2.	3.	2.	2.	2.	2.	1.	4.	4.	4.
K/RB	463.	895.	548.	342.	664.	520.	800.	736.	436.	537.	498.	529.	506.	448.	540.	887.	628.	701.	618.
K/SR	18.	14.	31.	25.	25.	20.	23.	21.	18.	29.	20.	35.	31.	19.	42.	28.	59.	52.	72.
BA/RB	14.2	13.0	7.7	10.7	10.8	10.5	14.5	14.2	10.0	10.0	33.8	11.2	11.2	11.0	9.5	17.0	6.4	11.1	8.5
RB/SR	0.038	0.015	0.056	0.072	0.038	0.039	0.029	0.029	0.040	0.053	0.040	0.067	0.061	0.043	0.078	0.032	0.095	0.074	0.114
SR87/86	.0	.0	.0	.0	.0	.0	.0	.7033	.0	.0	.0	.0	.0	.0	.7030	.0	.0	.7033	.0
TH	.31	.20	.61	1.0	.63	.61	.30	.65	.0	.89	.46	.81	1.4	.41	.99	.41	1.0	.99	.0
U	.22	.23	.62	.46	.39	.26	.26	.46	.0	.37	.29	.44	.46	1.2	.54	.24	.58	.55	.0
TH/U	1.4	0.9	1.0	2.2	1.6	2.3	1.2	1.4	0.0	2.4	1.6	1.8	3.1	0.3	1.8	1.7	1.9	1.8	0.0
K/U	39997.	29957.	13473.	19739.	20972.	37994.	25542.	14389.	0.	21763.	17788.	19691.	23613.	7471.	22398.	32513.	29055.	22942.	0.
HPU	0.5	0.4	0.8	0.8	0.6	0.6	0.4	0.6	0.0	0.7	0.4	0.7	0.9	1.2	0.9	0.5	1.0	0.9	0.0
LA	0.0	0.0	7.1	0.0	0.0	0.0	0.0	9.9	0.0	0.0	0.0	0.0	0.0	0.0	8.5	0.0	0.0	9.7	0.0
YB	.0	.0	2.4	.0	.0	.0	.0	4.4	.0	.0	.0	.0	.0	.0	3.1	.0	.0	3.5	.0
LA/YB	0.0	0.0	2.9	0.0	0.0	0.0	0.0	2.3	0.0	0.0	0.0	0.0	0.0	0.0	2.7	0.0	0.0	2.8	0.0
Y	23.	29.	29.	37.	34.	24.	45.	60.	25.	36.	72.	31.	40.	38.	42.	39.	68.	42.	54.
NI	0.	9.	17.	89.	4.	2.	6.	14.	12.	7.	10.	16.	6.	ND	ND	ND	ND	ND	ND
CO	13.	24.	25.	32.	19.	14.	20.	23.	12.	20.	24.	25.	15.	8.	16.	13.	12.	13.	6.
CR	3.	14.	31.	255.	5.	3.	8.	20.	2.	16.	23.	38.	8.	2.	4.	3.	4.	3.	3.
SC	11.	29.	27.	27.	22.	13.	25.	37.	12.	25.	28.	29.	16.	10.	22.	21.	19.	19.	17.
V	92.	254.	217.	198.	210.	108.	218.	223.	84.	225.	224.	235.	102.	56.	128.	132.	113.	90.	42.
CU	36.	34.	51.	62.	41.	19.	41.	41.	20.	47.	27.	28.	31.	6.	26.	12.	30.	32.	7.
NI/CO	0.0	0.4	0.7	2.8	0.2	0.1	0.3	0.6	1.0	0.4	0.4	0.6	0.4	0.0	0.0	0.0	0.0	0.0	0.0
V/NI	0.	28.	13.	2.	51.	54.	36.	16.	7.	32.	22.	14.	17.	0.	0.	0.	0.	0.	0.
MG/NI	0.	2741.	1621.	442.	3740.	9168.	3257.	1801.	1242.	3050.	2232.	1368.	2021.	0.	0.	0.	0.	0.	0.
SN	.0	.0	.42	.0	.0	.0	.0	.37	.0	.0	.0	.0	.0	.0	.52	.0	.0	.69	.0
HF	.0	.0	1.5	.0	.0	.0	.0	1.5	.0	.0	.0	.0	.0	.0	2.0	.0	.0	2.1	.0
ZR	53.	75.	100.	106.	108.	59.	83.	99.	59.	113.	99.	118.	146.	66.	134.	90.	150.	159.	226.
B	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.	0.
ZR/HF	0.	0.	66.	0.	0.	0.	0.	65.	0.	0.	0.	0.	0.	0.	66.	0.	0.	76.	0.
TI/ZR	71.	73.	58.	50.	54.	74.	65.	56.	71.	50.	56.	47.	37.	53.	45.	54.	36.	33.	26.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

* INDICATES SAMPLES FROM EASTERN GROUP

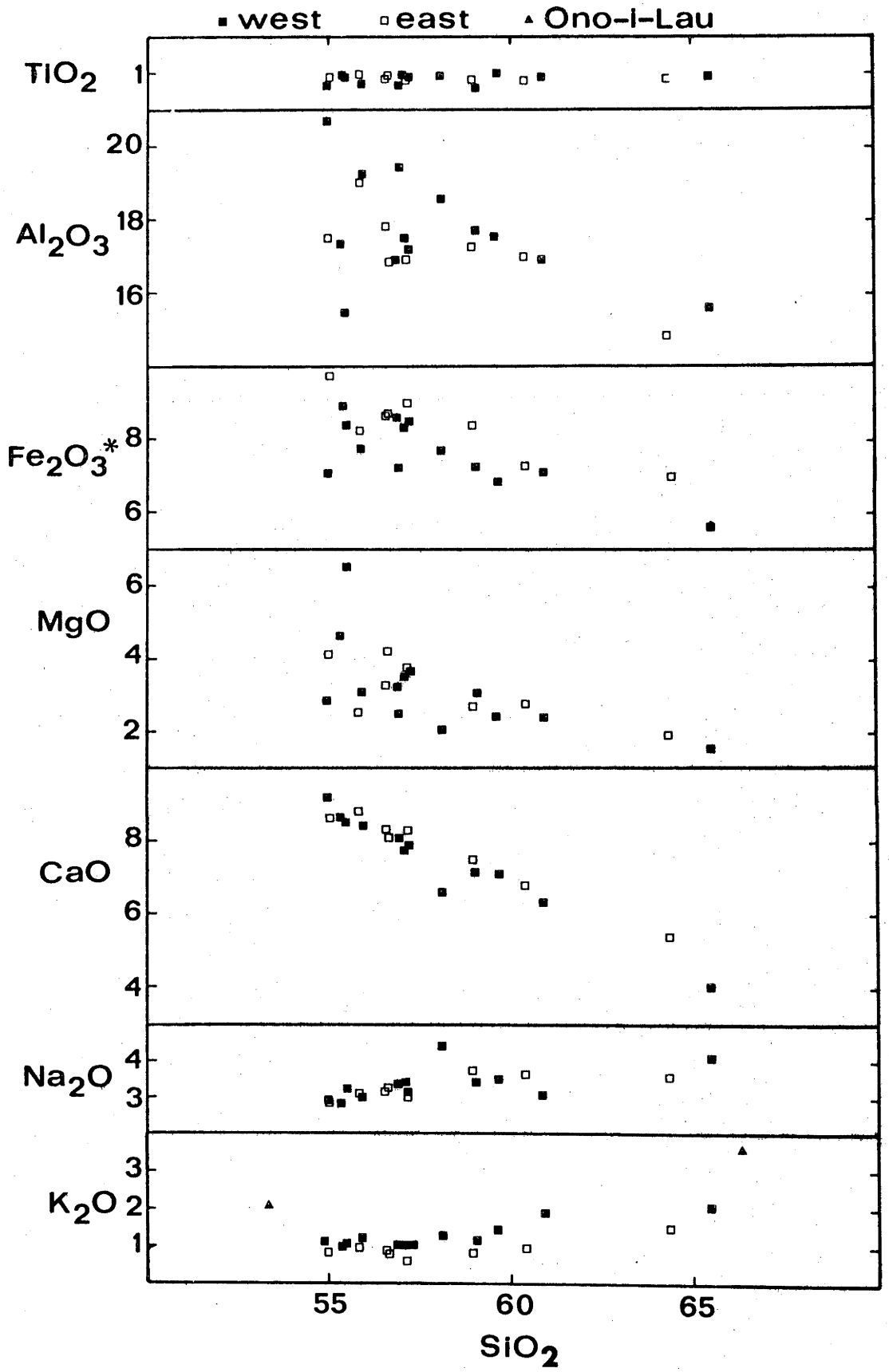

Figure 6.3 Major element oxides of Lau Volcanics. Data from Table 6.2 and Rodda (1969).

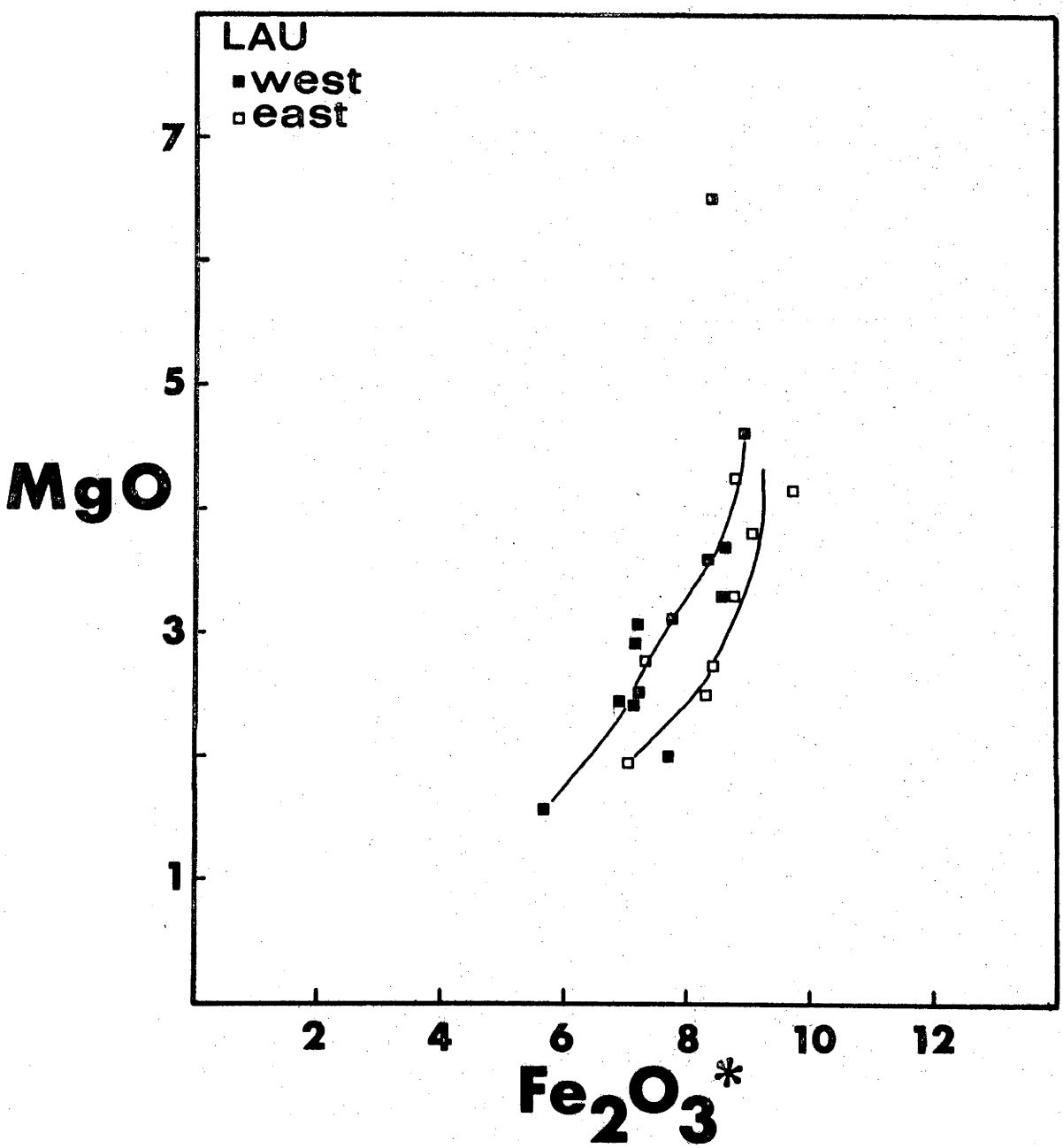

Figure 6.4 Fe-Mg variations in samples from eastern and western Lau Islands. Data sources as in Figure 6.3.

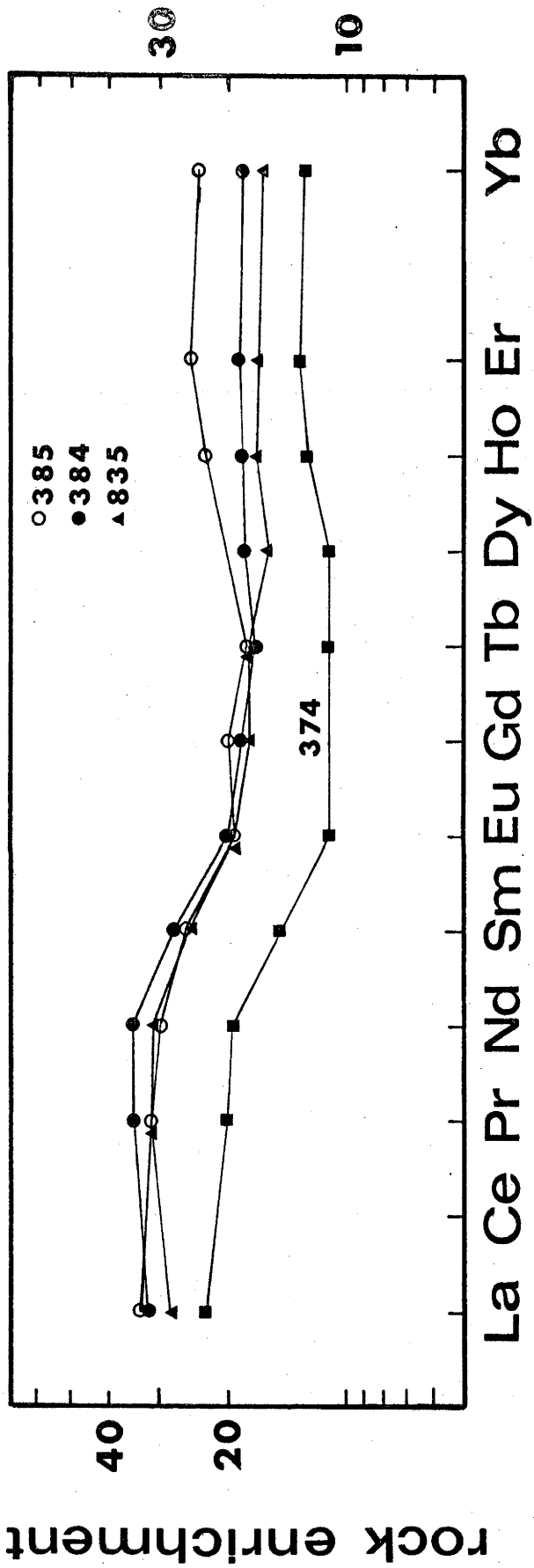

Figure 6.5 Rare earth patterns of Lau Volcanics.
Data from Table 6.3.

Table 6.3 Rare earth analyses of samples of the Lau Volcanics and from the Kadavu Group

	374	385	835	384	390
La	7.0	9.8	8.4	9.7	27.5
Ce	16.0	17.1	19.9	23.7	54.
Pr	3.0	3.7	3.7	4.2	8.4
Nd	14.3	17.8	17.2	20.2	32.
Sm	3.8	4.6	4.6	5.1	5.5
Eu	1.1	1.4	1.4	1.4	1.6
Gd	3.9	-	5.0	5.2	-
Tb	.60	.89	.86	.84	.62
Dy	3.7	6.2	4.8	5.1	3.4
Ho	.94	1.7	1.2	1.3	.58
Er	2.4	4.5	3.1	3.4	1.6
Yb	2.4	4.3	3.1	3.5	1.4

of two to three, slight enrichments of Pr and Nd, slightly negative Eu anomalies which do not exceed analytical uncertainties, and about 13 to 23 times chondritic heavy REE contents. Y concentrations are generally though less consistently higher in the east than west when rocks similarly fractionated from basaltic compositions are compared.

Two eastern group samples have 87Sr/86Sr ratios of 0.7033 with Rb/Sr ratios of 0.21 and 0.08; one western group sample has an 87Sr/86Sr ratio of 0.7030 and Rb/Sr of 0.22.

6.3 Korombasanga Volcanics: Pliocene

The Korombasanga Volcanics are olivine basalts named after Korobasaqa Hill, the highest point (248 m) on Vanua Balavu at the top of which limestone and not basalt (as implied by Ladd and Hoffmeister) crops out. Just north of Daliconi village they overlie Futuna Limestone and are therefore younger than Late Miocene in age. One K-Ar measurement (Table 6.1) suggests a Pliocene (3.5 ±0.2 m.y.) age. Ladd and Hoffmeister thought them subaerial and to also occur on Munia, Cikobia-i-Lau and Susui in the Exploring Islands and on Oneata to the south, largely because Alling found olivine in thin sections of rocks from these places. As mentioned above, I found none on Munia or Cikobia-i-Lau, but did find petrographically similar basalts in eastern Susui which I did not analyze.

Most of my samples are strongly porphyritic with large, zoned labradorite phenocrysts, 2 to 10% olivine, occasional clinopyroxene (0 to 10%), and an opaque. They are generally fresh with only olivine and minor glass alteration.

Five new analyses are given in Table 6.4 and an additional two are available from Foye (1918). Samples from near Korobasaqa Hill and adjacent headlands are hyp-normative olivine tholeiites. My sample (376) from the shores of Yanuyanu (opposite Lomaloma) is slightly ne-normative whereas Foye's is strongly hyp-normative (when $Fe_2O_3/FeO = 0.25$ for each); both have 2.4% TiO_2 but the analyses differ in SiO_2, MgO, Na_2O, K_2O, and P_2O_5. All have considerable Al_2O_3 (16 to 20%). The mainland samples have about 0.6% K, 10 ppm Rb, 420 ppm Sr, 160 ppm Ba, 2 ppm Pb, 0.2 ppm Th, 0.1 ppm U, 17 ppm Y, and 33 ppm Zr. The Yanuyanu sample has higher Rb, Ba, and Y and much higher Th, U, Ti, and Zr contents. Their 87Sr/86Sr ratios are analytically indistinguishable despite the difference in Rb/Sr ratios.

6.4 On odinites, hornblende-trachytes, and Ono-i-Lau

A. Odinites

Because some rocks from Vanua Balavu, Mago, and Kanbara have no phenocrysts other than orthopyroxene, Alling (op. cit., p. 201) identified them as "odinites", a name given to some German labradorite-augite-hornblende lamprophyres by Chelius in 1892. The Lauan rocks bear little resemblance in mineralogy (see Johannsen, 1937, vol. III, pp. 326-7) or composition (Washington, 1917, pp. 436 and 652) to these German or other rocks described as odinites. On Vanua Balavu, Ladd and Hoffmeister (1945, p. 27) consider these "odinites" part of the Korombasanga Volcanics; for Mago and Kabara they raise their status to that of an entirely different volcanic Group (the Mango Odinite), Quaternary in age. Their only evidence for distinction from the Lau Volcanics appears geomorphic (ibid., pp. 62-8 and 142-5). There is no evidence other than petrographic dissimilarity to distinguish them from Korombasanga Volcanics. Both the name and stratigraphic significance attached to these rocks seem misguided.

The analysis of one vesicular and altered sample, collected near Ladd and Hoffmeister's stations 53 and 54 on Vanua Balavu, is included in Table 6.4. It contains only pyroxene phenocrysts (15%), but most are clinopyroxene, although some orthopyroxene is present. The phenocrysts are fresh or have oxidized rims and are set in a matrix of

TABLE 6.4. KOROMBASANGA VOLCANICS AND MISCELLANY FROM LAU.

DATA	378	376	368	372	831	371	382	393
SIO2	47.51	47.98	48.10	48.15	50.51	60.22	48.35	66.35
TIO2	0.87	2.45	0.90	0.84	0.92	0.56	0.70	0.98
AL203	18.36	16.77	19.66	18.32	18.78	18.06	14.11	18.16
FE203	11.60	11.92	10.72	10.50	10.78	6.65	10.19	2.12
FEO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
MNO	0.20	0.19	0.18	0.18	0.20	0.20	0.17	0.02
MGO	6.78	6.84	5.32	7.13	5.43	2.62	11.87	0.18
CAO	11.72	8.55	11.90	12.15	10.17	7.35	11.74	3.24
NA2O	1.99	3.64	2.24	1.92	2.37	3.07	1.91	4.90
K2O	0.77	1.16	0.74	0.62	0.61	1.01	0.63	3.72
P2O5	0.19	0.50	0.23	0.19	0.22	0.26	0.33	0.33
(LOI)	1.41	0.98	0.60	0.64	1.24	1.59	4.24	1.40
(TOTAL)	100.91	99.80	100.02	99.80	98.88	99.42	100.58	99.54
NA2O/K2O	2.6	3.1	3.0	3.1	3.8	3.0	3.0	1.3
MG/MG+FE2	0.59	0.58	0.55	0.62	0.55	0.49	0.74	0.17
%K	0.64	0.97	0.62	0.51	0.51	0.84	0.53	3.09
RB	13.	21.	8.5	11.	8.7	20.	8.3	86.
SR	421.	453.	436.	381.	452.	504.	385.	338.
BA	161.	233.	171.	131.	184.	204.	197.	438.
PB	2.	2.	1.	2.	2.	1.	2.	6.
K/RB	492.	455.	723.	460.	586.	411.	633.	361.
K/SR	15.	21.	14.	13.	11.	17.	14.	91.
BA/RB	12.3	11.0	20.0	11.8	21.2	10.0	23.7	5.1
RB/SR	0.031	0.047	0.020	0.029	0.019	0.041	0.022	0.253
SR87/86	.0	.7034	.0	.7035	.0	.0	.0	.0
TH	.20	2.9	.20	.20	.20	.51	.0	4.2
U	.10	.82	.10	.10	.10	.20	.0	1.5
TH/U	2.0	3.6	2.0	2.0	2.0	2.5	0.0	2.7
K/U	63920.	11786.	61430.	50638.	49808.	41091.	0.	20200.
HPU	0.3	1.4	0.3	0.3	0.3	0.5	0.0	2.8
LA	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
YB	.0	.0	.0	.0	.0	.0	.0	.0
LA/YB	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Y	16.	31.	18.	15.	19.	22.	13.	51.
NI	22.	81.	14.	35.	17.	10.	187.	0.
CO	38.	35.	31.	35.	30.	9.	47.	0.
CR	47.	111.	16.	101.	35.	3.	737.	2.
SC	28.	18.	18.	33.	27.	10.	39.	15.
V	276.	162.	144.	252.	264.	71.	239.	25.
CU	70.	45.	83.	86.	42.	8.	104.	5.
NI/CO	0.6	2.3	0.5	1.0	0.6	1.1	4.0	0.0
V/NI	13.	2.	10.	7.	15.	7.	1.	0.
MG/NI	1851.	510.	2279.	1218.	1880.	1544.	383.	0.
SN	.0	.0	.0	.0	.0	.0	.0	.0
HF	.0	.0	.0	.0	.0	.0	.0	.0
ZR	29.	225.	35.	32.	37.	52.	34.	214.
B	0.	0.	0.	0.	0.	0.	0.	0.
ZR/HF	0.	0.	0.	0.	0.	0.	0.	0.
TI/ZR	180.	65.	152.	155.	150.	65.	122.	27.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

371 = Hb-trachyte; 382 = "Odinite"; 393 = Ono-i-Lau sample

devitrified glass, andesine, pyroxene, and opaques. It is similar yet dissimilar to analyses of both Lau and Korombasanga Volcanics. The odinites' field association with olivine-bearing rocks in Vanua Balavu and Kabara (see Foye, 1918) suggests a genetic correlation with Korombasanga Volcanics in which olivine reacts with olivine-tholeiitic liquid to yield orthopyroxene. There is, however, no petrographic evidence for this reaction in Korombasanga basalts and no petrologic evidence with which to dismiss plagioclase from the liquidus by such reaction. An alternative is to regard "odinites" as cumulates from Lau Volcanic magmas which, at times, precipitated only plagioclase and orthopyroxene. This seems more reasonable to me, especially in light of the groundmass andesine.

B. Hornblende-trachytes

Hornblende-bearing trachytes and andesites are thought to overlie Futuna Limestone on Vanua Balavu (Ladd and Hoffmeister, 1945, pp. 28 and 31), and are therefore included with the Korombasanga Volcanics. The analysis of one sample, collected near their station 94, is given in Table 6.4. It is andesite whereas all other analyzed Korombasanga rocks are basalts, so definitive comparison is impossible. It differs, however, in composition as well as texture and mineralogy from Lau Volcanic andesites, especially in K, Na, Th, U, Y, Ni and Zr contents. It is similar in composition to the hornblende-andesites that are spatially contiguous with shoshonites and basalts in Viti Levu (see section 4.4E), although containing less K, Na, Rb, Sr, Ba, Pb, Th, and U just as Korombasanga basalts have lower concentrations of these elements than do basic rocks of northern Viti Levu.

Field relationship to the Futuna Limestone rather than chemical composition will decide their stratigraphic position, especially because hornblende-gabbro xenoliths can be found in Lau Volcanics (see section 6.2B).

C. Ono-i-Lau

Ono-i-Lau is the southernmost inhabited Fiji island and little has been published concerning its geology. The most recent account appears in the FGSD Annual Report for 1969 (p. 9) in connection with a groundwater survey. Its description as being "composed entirely of andesite agglomerate and dikes" and at most 100 m high (ibid.) suggest correlation with the Lau Volcanics, although Foye's (1918) report that olivine and andesine are common phenocrysts does not. Foye includes

one analysis of a basaltic-andesite and I present another of a dacite in Table 6.4, the rock having been collected by R. Richmond during the above-mentioned groundwater study. Both analyses are notable for their high Ti, Na, and K values. The latter also has much higher Rb, Ba, Pb, Th, U, and Zr contents than Lau Volcanic dacites from the Exploring Islands. One K-Ar measurement gave 5.9 ±0.1 m.y. as the dacite's age (Table 6.1). The rock contains at most 15% plagioclase phenocrysts, usually fresh labradorite, set in a matrix of devitrifying glass, plagioclase microlites with flow orientations, specks of pyroxene, and opaques. The glass could have leaked Ar resulting in an underestimate of age.

The relationship of these three single samples and the rock types or areas they represent obviously remains undecided.

Chapter 7. Taveuni, the Lomaiviti Islands, Kadavu, and the Mamanucu and Yasawa Islands

By early 1972 little had been published concerning the geology of these islands. Of them, I visited only Taveuni and collected samples from locations indicated in Figure 5.1. Other samples were donations and are identified by island in Table 7.2, and described geographically and petrographically in Appendix 2.

7.1 Taveuni

Taveuni, sited by Tasman in 1643 but settled at least 2000 years before, is the third largest Fiji island (420 km^2). Its northwest-trending axial ridge (~1230 m high) includes some 120 volcanic centers (Rickard, 1966) along a well-defined fissure zone from which have come recent basalt flows which blanket the island and provide its modern geomorphic features. Some vents were active between 710 ±80 and 2050 ±150 years ago as these are C14 dates of charcoal collected from inhabited sites above and below volcanic ash at a southern Taveuni archaeological excavation (E. Frost, written comm., 1970). The island is now being geologically mapped by W. Hindle, supplementing Rickard's and Ibbotson's (1960b) brief accounts.

Taveuni rocks are very fresh and frequently vesicular basalts with 5 to 15% labradorite, 2 to 10% olivine, and less than 5% clinopyroxene phenocrysts. These same minerals plus an opaque form a crystalline groundmass.

Seven new analyses, possibly all from the same, most recent eruption(s), are listed in Table 7.1. One other is available from Rodda (1969) and two from Hindle (1970). All have 1.6 to 2.1% TiO_2 and are ne-normative alkali olivine basalts. Most have $Mg/Mg+Fe^{+2}$ ratios >0.65 although no ultramafic xenoliths are reported. My analyses are quite uniform with coefficients of variation exceeding 15% only for MgO, Ni, Cr, Th, and U. Th and U variability could reflect secular disequilibrium as all samples may not be older than 1 m.y. (see Appendix 1, section 3.B). Some average element ratios are: K/Rb = 384; Rb/Sr = 0.046; Th/U = 2.8; K/U = 9000; Ni/Co = 2.5; and V/Ni = 4. One 87Sr/86Sr ratio is 0.7035.

Also in Table 7.1 is an analysis of what appeared to be black glass with chonchoidal fracture filling a 1 to 2 cm^3 vesicle in sample 824. It is similar in composition to Taveuni basalts except for being low in Al_2O_3, high in CaO and MgO, bereft of alkalis, and high in Sc and V.

TABLE 7.1. TAVEUNI VOLCANICS.

DATA	827	829	824	828	830	825	826	P824
SIO2	47.17	47.80	47.85	48.02	48.98	48.90	49.16	47.32
TIO2	1.62	1.83	1.76	1.90	1.82	1.79	1.72	2.03
AL203	16.23	16.89	16.48	16.97	18.89	18.08	18.95	7.17
FE203	10.21	9.57	10.16	10.42	8.95	10.34	10.00	10.30
FEO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
MNO	0.17	0.16	0.17	0.17	0.15	0.17	0.17	0.20
MGO	9.44	8.86	8.33	7.43	5.34	5.22	5.09	13.29
CAO	11.08	10.16	10.61	10.32	11.11	10.27	9.87	19.57
NA2O	2.81	3.33	3.12	3.15	3.43	3.76	3.66	0.0
K2O	0.88	0.95	1.09	1.10	0.91	0.98	0.94	0.06
P2O5	0.39	0.43	0.44	0.53	0.42	0.48	0.45	0.05
(LOI)	0.0	0.0	0.0	0.31	0.37	0.0	0.0	0.0
(TOTAL)	100.08	99.87	100.02	99.19	99.68	99.19	99.43	98.62
NA2O/K2O	3.2	3.5	2.9	2.9	3.8	3.8	3.9	0.0
MG/MG+FE2	0.69	0.69	0.67	0.63	0.59	0.55	0.55	0.76
%K	0.73	0.79	0.90	0.92	0.75	0.81	0.78	0.05
RB	21.	21.	27.	24.	20.	18.	18.	.0
SR	386.	466.	420.	526.	495.	516.	529.	72.
BA	290.	276.	246.	356.	272.	299.	290.	30.
PB	3.	2.	3.	2.	2.	3.	2.	0.
K/RB	348.	376.	335.	377.	374.	447.	429.	0.
K/SR	19.	17.	22.	17.	15.	16.	15.	7.
BA/RB	13.8	13.1	9.1	14.7	13.5	16.5	16.0	0.0
RB/SR	0.054	0.045	0.064	0.046	0.041	0.035	0.034	0.0
SR87/86	.0	.0	.0	.0	.0	.0	.7035	.0
TH	2.8	2.7	2.9	3.2	2.6	1.7	1.6	.0
U	1.4	.67	1.8	.87	.56	.95	.91	.0
TH/U	2.0	4.0	1.6	3.7	4.6	1.8	1.8	0.0
K/U	5218.	11771.	5027.	10521.	13341.	8566.	8578.	0.
HPU	1.8	1.2	2.1	1.5	1.1	1.3	1.2	0.0
LA	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
YB	.0	.0	.0	.0	.0	.0	.0	.0
LA/YB	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Y	30.	33.	33.	34.	36.	35.	33.	33.
NI	177.	170.	149.	120.	59.	28.	28.	61.
CO	49.	43.	44.	44.	31.	34.	35.	48.
CR	500.	403.	441.	242.	164.	30.	40.	58.
SC	27.	30.	27.	35.	25.	26.	26.	122.
V	220.	189.	232.	211.	230.	267.	264.	649.
CU	67.	57.	69.	74.	45.	58.	61.	69.
NI/CO	3.6	4.0	3.4	2.7	1.9	0.8	0.8	1.3
V/NI	1.	1.	2.	2.	4.	9.	9.	11.
MG/NI	322.	314.	337.	373.	542.	1116.	1090.	1318.
SN	.0	.0	.0	.0	.0	.0	.0	.0
HF	.0	.0	.0	.0	.0	.0	.0	.0
ZR	125.	161.	142.	163.	150.	137.	123.	0.
B	0.	0.	0.	0.	0.	0.	0.	0.
ZR/HF	0.	0.	0.	0.	0.	0.	0.	0.
TI/ZR	78.	68.	74.	70.	73.	78.	84.	0.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

P824 INDICATES CLINOPYROXENE FROM SAMPLE 824

An X-ray powder photograph showed it to be pyroxene, not glass. As pyroxene it has the composition of augite ($Mg_{41}Ca_{43}Fe_{16}$) with the following structural formula (calculated without Na_2O or Fe_2O_3): Si = 1.84, Al = .16 (Σ=2.00); Al = .06, Ti = .06, Fe^{+2} = .30, Mn = .07, Mg = .77, Ca = .81 (Σ=2.07).

7.2 The Lomaiviti Islands

Lomaiviti or Central Fiji (Figure 1.1) is a group of islands including Ovalau, Koro, Makogai, Wakaya, Gau, Totoya, Matuku, Batiki, Nairai, and Moala. My samples, collected by P. Rodda and F. Coulson, are from the last four of these.

Lomaiviti geology has recently been mapped by F. Coulson (in prep.), although I have seen only his summaries in FGSD Annual Reports for 1968 and 1969. Ibbotson (1961) mapped Ovalau. According to these sources, the islands are geomorphically young and expose mostly olivine-basalts with plagioclase, augite, and olivine, singly and in various combinations, as dominant phenocrysts. Most have Pliocene K-Ar ages (Coulson, pers. comm., 1971). Spinel-lherzolite nodules were found on Koro which may be the youngest of the group. Hornblende-bearing rocks ("andesites" in field descriptions) are reported from Ovalau and nearby islets (Ibbotson, 1961) and Nairai (Coulson in FGSD Annual Report for 1969, p. 6).

Six new analyses, three for rocks from or near Nairai and one each from Batiki, Moala, and Matuku, are given in Table 7.2; Hindle (1970) reported one each from Nairai, Vatu-i-Cake, and Makogai. Those from Nairai include one transitional olivine basalt (1.3% normative ne when Fe_2O_3/FeO = 0.25) and three qz-normative basaltic-andesites which are the "andesites" mentioned above. Rb, Ba, Pb, Th, and U contents are low in the basalt but more than doubled in the basaltic-andesites. One of the latter has an 87Sr/86Sr ratio of 0.7034. One partial analysis of a laminated, tuffaceous "andesinite" from Naigani near Ovalau (see FGSD Annual Report for 1969, p.6) has 70% SiO_2 and 17% Al_2O_3 but only 1.3% K_2O.

When Fe_2O_3/FeO = 0.25, my samples from Moala, Batiki, and Matuku are transitional olivine basalts (0.3% normative hyp to 1.7% normative ne). Hindles's sample from Vatu-i-Cake is more strongly ne-normative whereas his from Makogai is more strongly hyp-normative. All have <1% TiO_2 and <100 ppm Zr. The Moala, Batiki, and Matuku basalts, although little more undersaturated than the Nairai sample in normative

TABLE 7.2 LOMAIVITI, KADAVU, MAMANUCU, AND YASAWA ISLANDS.

	LOMAIVITI						KADAVU	MAMANUCU		YASAWA
		Nairai		Moala	Batiki	Matuku	Ono	Yavuriba	Tavua	"White Rock"
DATA	842	841	843	394	840	389	390	392	391	844
SIO2	46.68	53.30	54.43	48.74	48.85	49.61	60.64	57.98	61.48	66.19
TIO2	1.12	0.76	0.77	0.70	0.86	0.92	0.62	0.80	0.59	0.35
AL2O3	16.14	19.42	18.01	13.14	14.31	17.21	17.08	16.82	16.80	16.64
FE2O3	12.34	9.38	8.70	11.17	10.73	9.96	5.92	7.92	6.63	4.22
FEO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
MNO	0.21	0.20	0.25	0.20	0.19	0.16	0.11	0.16	0.17	0.06
MGO	7.55	3.77	3.54	11.03	8.77	5.38	2.18	3.61	2.52	1.71
CAO	13.08	8.90	8.59	11.30	12.61	11.95	5.74	7.79	4.72	3.98
NA2O	2.12	3.06	4.18	1.84	2.07	2.50	4.47	3.46	5.01	3.87
K2O	0.55	0.74	1.16	1.59	1.36	1.94	2.89	1.25	1.83	2.73
P2O5	0.22	0.47	0.37	0.29	0.25	0.35	0.34	0.22	0.23	0.25
(LOI)	1.92	1.56	2.58	2.82	1.64	3.83	0.88	2.10	1.83	2.32
(TOTAL)	98.95	99.16	99.70	99.86	99.22	99.69	99.49	99.69	100.05	99.15
NA2O/K2O	3.9	4.2	3.6	1.2	1.5	1.3	1.5	2.8	2.7	1.4
MG/MG+FE2	0.60	0.49	0.50	0.71	0.66	0.57	0.47	0.52	0.48	0.50
%K	0.45	0.61	0.97	1.32	1.13	1.61	2.40	1.04	1.52	2.26
RB	9.9	18.	22.	32.	24.	38.	52.	40.	32.	61.
SR	362.	525.	510.	445.	478.	752.	1349.	376.	377.	1299.
BA	127.	368.	332.	330.	512.	386.	847.	269.	292.	1281.
PB	1.	3.	4.	2.	3.	4.	8.	3.	2.	10.
K/RB	458.	332.	447.	412.	480.	429.	464.	260.	482.	371.
K/SR	13.	12.	19.	30.	24.	21.	18.	28.	40.	17.
BA/RB	12.8	19.9	15.3	10.3	21.7	10.3	16.4	6.7	9.3	21.0
RB/SR	0.027	0.035	0.042	0.072	0.049	0.050	0.038	0.106	0.084	0.047
SR87/86	.0	.7034	.0	.7038	.0	.0	.7032	.0	.0	.0
TH	.41	1.7	1.5	.41	.31	.52	4.3	1.4	1.2	2.3
U	.21	.47	.62	.25	.32	.26	1.7	.0	.53	1.4
TH/U	2.0	3.7	2.5	1.7	1.0	2.0	2.5	0.0	2.3	1.6
K/U	21998.	12993.	15634.	53267.	35615.	61762.	13917.	0.	28735.	15654.
HPU	0.4	0.9	1.0	0.6	0.6	0.7	2.8	0.0	1.0	2.1
LA	0.0	0.0	0.0	0.0	0.0	0.0	27.9	0.0	0.0	0.0
YB	.0	.0	.0	.0	.0	.0	1.4	.0	.0	.0
LA/YB	0.0	0.0	0.0	0.0	0.0	0.0	19.6	0.0	0.0	0.0
Y	25.	32.	34.	20.	20.	22.	17.	29.	26.	7.
NI	110.	11.	67.	155.	45.	58.	6.	11.	4.	8.
CO	52.	25.	28.	52.	43.	33.	18.	22.	13.	11.
CR	340.	5.	137.	376.	225.	106.	8.	26.	9.	44.
SC	40.	11.	12.	39.	40.	25.	12.	25.	11.	11.
V	269.	133.	126.	246.	272.	265.	168.	156.	82.	105.
CU	92.	36.	24.	180.	134.	121.	60.	31.	15.	35.
NI/CO	2.1	0.5	2.4	3.0	1.0	1.8	0.3	0.5	0.3	0.7
V/NI	2.	12.	2.	2.	6.	5.	28.	14.	20.	13.
MG/NI	413.	2018.	319.	430.	1173.	556.	2161.	1930.	3740.	1252.
SN	.0	.0	.0	.0	.0	.0	1.8	.0	.0	.0
HF	.0	.0	.0	.0	.0	.0	2.3	.0	.0	.0
ZR	53.	112.	109.	39.	42.	49.	136.	124.	132.	63.
B	0.	0.	11.	0.	6.	0.	8.	0.	0.	12.
ZR/HF	0.	0.	0.	0.	0.	0.	58.	0.	0.	0.
TI/ZR	128.	41.	42.	107.	123.	112.	27.	39.	27.	33.

ND = NOT DETECTABLE; 0. = NOT DETERMINED

Figure 7.1 Rare earth pattern of a sample from the Kadavu Group. Data from Table 6.3.

mineralogy,have considerably higher concentrations of K, Rb, Ba, and Sr; lower Na_2O/K_2O ratios (1.2 to 1.5 rather than 3.9); and higher Rb/Sr, K/Sr, K/U, and 87Sr/86Sr ratios. In these respects they are similar to basalts from the neighboring Rakiraki area of Viti Levu (section 4.4D).

Coulson (pers. comm., 1971) has found the lherzolite-bearing basalts of Koro to be ne-normative with about 2% TiO_2, like those of Taveuni.

7.3 Kadavu

The Kadavu group includes the fourth largest Fiji island (Kadavu: 125 km^2) and Ono, where my sample 390 was collected by P. Rodda. Kadavu has peaks ∿700 m high and is separated from Viti Levu by water 2 to 2.5 km deep. Robertson (1965) reports one Bouguer anomaly of 110 mgal which suggests crust (σ = 2.67 g/cm^3) about 25 km thick.

Little is known of Kadavu's geology and nothing of its age. Renard (1889), describing rocks returned by the Challenger expedition, noted mostly hornblende-andesites containing labradorite, hornblende, biotite, magnetite, and occassional augite and hypersthene. The amphiboles invariably have reaction rims of augite and magnetite. Augite phenocrysts, when more abundant than hornblende, are also associated with a glassier groundmass. Foye (1918) corroborates Renard, adding olivine-andesites from Kadavu and olivine-basalts from Dravuni to the list of observed petrographic rock types. On the basis of geomorphology and these rock types, Phillips (1964) considered Kadavu Late Miocene in age.

Foye provides one analysis of a rock from northern Kadavu to which I add another from Ono (Table 7.2). Both samples are dominated by plagioclase and altered hornblende phenocrysts with minor pyroxenes and oxides. Their analyses are similar except in MgO and CaO; both have high K_2O contents and Na_2O/K_2O ratios around 1.6. Large cation trace elements are concentrated in sample 390 including the light REE (La/Yb = 20: Figure 7.1 and Table 6.3). So also are some chalcophile elements like Sb and Sn. Despite high alkali contents the 87/86 ratio of its Sr is 0.7032.

7.4 The Mamanucu and Yasawa Islands

These islands lie along the westernmost edge of the Lau-Colville Ridge, northwest of Viti Levu (Figure 1.1). P. Rodda is currently mapping them and an early report of his observations appears in the

FGSD Annual Report for 1969. Previous descriptions include those of Foye (1918), Smith and Dickinson (1965), and Dickinson (1966). Although no explicit evidence of age is known, the submarine character of volcanism (pillow lavas with matrices of red calcareous ooze), presence of over-turned folds, and prevalence of zeolite to greenschist facies assemblages, suggest correlation with first period rocks of Viti Levu. Dickinson's and Foye's descriptions of frequent green and brown hornblende and orthopyroxene phenocrysts in Mamanucu and Yasawa rocks, however, contrast with what is known of first period Viti Levu material (section 4.2B).

Foye presents one analysis, and I another three in Table 7.2 for samples collected by P. Rodda. All contain 5 to 20% plagioclase, both andesites have orthopyroxene, and the more acid andesite and dacite have hornblende phenocrysts. Both andesites contain gabbroic inclusions having the same mineralogy as their respective phenocrysts: norite in sample 392, hornblende-gabbro in 391.

They vary considerably but all three are closer in composition as well as mineralogy to Namosi and Kadavu rocks than to most known from the Wainimala Group. Sample 844, from a solitary and reef-less dacite plug, has only 7 ppm Y yet high concentrations of most large cations, and could have a very high La/Yb ratio.

Chapter 8. The Evolution of Fiji

8.1 Tectonic implications of Fijian geochemistry

As discussed in sections 2.1 and 2.2, Fiji now occupies an anomalous tectonic position. Its previous relationship to the plates and plate boundaries around it are unclear yet integral to an understanding of how Fiji developed. Did it arise within an oceanic plate as an island like Hawaii? Did it evolve at a plate boundary above a subduction zone like Tonga? Or did it develop uniquely, always caught within the complex boundary between Pacific and Australian plates and never with a simple relationship of any kind to crustal subduction?

Geology has few ways to answer such questions. Certain sedimentary and metamorphic rocks have come to be associated with a subduction environment (Dickinson, 1971c) but are absent from Fiji. Only the characteristics of Fiji's igneous rocks remain as useful evidence. Of these, three are singularly important. First, pre-Pliocene rocks are mostly basaltic-andesite and andesite in composition with <1.2% TiO_2. All modern volcanic provinces for which this is true are associated with a subduction environment. Fiji, by uniformitarian inference, was also. This is the modern significance of Fiji lying west of Marshall's (1912) "andesite line".

Second, Late Miocene volcanic rocks, which constitute most of those presently exposed in Fiji, display regional variations in composition similar to those documented in chapter 3 and consistent with conventional northeast to southwest subduction along the north and east edges of the Lau-Colville Ridge. These transverse variations and the linearity of the ridge as a physiographic feature both suggest that Fiji was once part of a <u>normal</u> island arc subduction environment.

Third, magma composition changed dramatically throughout Fiji at about the beginning of the Pliocene, 5 to 6 m.y. ago. This change and the inference that most of the adjacent Fiji Plateau and Lau Basin are no older than Pliocene in age (section 2.2), suggest that Fiji's enigmatic characteristics are recent acquisitions.

I have presented these ideas more fully elsewhere (Gill and Gorton, Appendix 3) and now wish to examine the assumptions and requirements of and evidence for my proposal in greater detail.

8.2 Geochemical evidence for the proposal

The first period of Viti Levu volcanism (section 4.2) yielded rocks with unambiguously intermediate composition. No tectonic polarity can be inferred from the Fijian record, but there may not have been any northeast-dipping subduction in eastern Melanesia at this time (section 2.3).

Second period Fijian volcanism was also predominantly intermediate in composition. Its products include the Mathuandrove Super-Group of Vanua Levu (section 5.2) and the Namosi Andesites of Viti Levu (section 4.3), probably included the rocks of the Kadavu Group (section 7.3) and the Lau Volcanics (section 6.2), and possibly included some of those from the Mamanucu Islands (section 7.4). Contact between first and second period volcanics is exposed only on Viti Levu where it is always unconformable. The geochemistry of these Late Miocene volcanic rocks is summarized in Table 8.1. Comparison with the data of Table 3.1 and Figures 3.1 to 3.3 suggest north to south subduction beneath Vanua Levu, Viti Levu, and Kadavu, and east to west subduction beneath Lau during this period. (Quantitative estimates of the dip of this paleoseismic zone using Dickinson's (1970a) K-h plots and reasoning are unconvincing.)

The third period of Fijian volcanism was dominantly though not entirely basaltic or shoshonitic. Its products include rocks from several centers on Viti Levu (section 4.4), the Mbua Volcanic Group and related olivine-bearing intrusives of Vanua Levu (section 5.3), the Korombasanga Volcanics of Lau (section 6.3), and the rocks of Taveuni and Lomaiviti (sections 7.1 and 7.2). The geochemistry of these Pliocene to Recent rocks is summarized in Table 8.2, and discussed more fully in chapter 12. Temporal variations in their compositions may reflect effects within the upper mantle of a migrating subduction zone, but my data are inadequate to evaluate this possibility properly.

8.3 Assumptions and requirements of the proposal

The interpretation summarized above is not unique, but it appears the simplest. For it to be correct the following conditions must be met.

First, the analytical data must be reliable and reflect magma compositions, not post-magmatic events. Reliability can be assessed from information provided in Appendix 1. Potassium contents were cross-checked by three independent techniques. Of the polarity-significant elements (i.e. those listed in Figure 3.1), only Ba and very low Rb, Th, and U contents are expected to be less accurate than $\pm$10%

Table 8.1. Comparison of average late Miocene basaltic-andesite and andesite compositions. Arrow indicates inferred direction of underthrusting.

	Vanua Levu		Viti Levu		Kadavu	Lau			
						EAST		WEST	
	→		→		→	→		→	
	$\bar{x}$	σ	$\bar{x}$	σ	$\bar{x}$	$\bar{x}$	σ	$\bar{x}$	σ
(N)	(11)		(8)		(1)	(7)		(12)	
SiO_2	56.6	4.7	58.4	3.4	60.6	58.0	3.2	58.1	3.2
TiO_2	0.8	0.2	0.7	0.1	0.6	0.9	.05	0.8	0.1
Al_2O_3	17.3	1.2	17.4	0.5	17.1	17.1	1.3	17.8	1.5
Fe_2O_3*	8.5	2.2	7.0	1.3	5.9	8.4	1.0	7.6	0.9
MgO	4.2	2.0	3.4	0.8	2.2	3.2	0.8	3.1	1.5
CaO	8.7	1.8	7.5	1.1	5.7	7.8	1.2	7.5	1.4
Na_2O	2.9	0.5	3.9	0.4	4.5	3.2	0.3	3.4	0.5
K_2O	.69	.24	1.23	.14	2.9	.93	.29	1.28	.38
P_2O_5	.17	.05	.22	.04	.34	.19	.04	.29	.08
Rb	11.	6.	21.	7.	52.	10.	3.	20.	4.
Ba	230.	76.	390.	50.	85.0	170.	80.	200.	48.
Sr	370.	170.	540.	70.	1350.	310.	90.	350.	110.
Pb	2.	1.	4.	1.	8.	2.	1.	2.	1.
Th	.47	.21	1.7	0.2	4.3	.51	.26	.66	.43
U	.23	.03	.64	0.1	1.7	.34	.12	.42	.32
Yb	2.1	-	1.4	-	1.4	4.0		2.8	-
Y	33.	27.	22.	3.	17.	45.	15.	36.	12.
K/Rb	630.	290.	520.	130.	460.	740.	140.	510.	80.
Ba/Rb	26.	14.	20.	6.	16.	16.	8.	10.	2.
Rb/Sr	.031	.017	.038	.013	.038	.037	.018	.063	.02
$^{87}/^{86}Sr$	.7040	-	.7037	-	.7032	-	-	-	-
Na/K	4.7	1.9	3.2	0.3	1.5	3.7	0.8	2.8	0.5
Th/U	2.0	0.7	2.7	0.4	2.5	1.4	0.3	1.5	1.0
K/U	25K	8K	16K	2K	14K	23K	6K	20K	13K
Pb/U	9.	-	6.	-	5.	4.	-	3.	-
La/Yb	4.	-	7.	-	20.	2.6	-	2.8	-

*Total Fe as Fe_2O_3

Table 8.2. Comparison of Pliocene to Recent rocks having <52% SiO_2

	Sabeto	Tavua			Rakiraki	Nakobalevu	Bua	Korobasaga		Taveuni	Nairai	Other Lomaiviti
		Absaro-kite	Shosho-nite	Bana-kite				A.	B.			
N	15	8	3	1	5	2	3	4	1	10	1	3
>2%qz	-	-	-	-	-	-	-	-	-	-	-	-
<2%qz;>2%hyp	2	1	-	-	-	2	3	4	-	-	-	1
<2%hyp;<2%ne	5	1	2	1	3	-	-	-	1	4	1	2
>2%ne	8	6	1	-	2	-	-	-	-	6	-	-
SiO_2	48-53	48-52	52-54	56-58	47-50	50.	46-50	47-50	48.	46-49	47.	48-50
Max. $Mg/Mg+Fe^{+2}$	0.61	>0.7	.52	.49	>0.7	.62	.62	.63	.63	>0.7	.61	>0.7
Al_2O_3	15-20	12-18	16-21	18-20	11-20	18.	16-18	18-19	17.	16-19	16.	11-17
TiO_2	0.7	0.65	0.65	0.55	0.8	0.8	2.4	0.9	2.5	1.8	1.1	0.6-0.9
Na_2O/K_2O	1.0	0.9	1.0	0.6	1.4	2.8	3.6	3.1	3.1	3.4	3.9	0.9-1.6
K_2O	3%	3%	4%	6%	1.8%	0.8%	0.8%	0.7%	1.2%	1.0%	0.5%	1.1-2.2%
ppm Rb	60.	70.	100.	170.	40.	13.	14.	10.	21.	21.	10.	24-50.
Sr	1300.	1100.	1450.	1300.	880.	500.	510.	420.	450.	475.	360.	450-1050.
Ba	625.	780.	860.	450.	550.	150.	230.	160.	230.	290.	130.	330-520.
Pb	8.	7.	12.	16.	6.	2.	3.	2.	2.	2.	1.	2-4.
K/Rb	490.	430.	400.	300.	420.	550.	540.	560.	450.	385.	460.	400-500.
Rb/Sr	0.044	.056	.069	.126	.045	.026	.029	.025	.047	.046	.027	.05-.07
Ba/Rb	11.	13.	7.	5.	15.	12.	17.	16.	11.	14.	13.	10-22.
K/Sr	21.	22.	25.	38.	18.	15.	15.	13.	21.	17.	13.	21-30.
87Sr/86Sr	0.7041	.7038	.7039	.7039	.7040	.7032	.7041	.7035	.7034	.7035	.7034**	.7039
Th	1.5	1.0	1.6	3.5	1.2	0.8	2.0	0.2	2.9	2.5	0.4	0.3-.5
U	0.8	0.5	0.6	1.7	0.4	0.4	0.4	0.1	0.8	1.0	0.2	0.1-.3
Th/U	2.0	2.2	2.5	2.1	3.0	2.0	4.5	2.	3.5	2.5	2.0	1-5.
K/U	46K	58K	50K	30K	38K	19K	16K	56K	12K	9K	22K	>36K
Y	19.	22.	26.	27.	21.	19.	29.	17.	31.	33.	25.	20-22.
Zr	48.	35.	67.	97.	61.	62.	170.	33.	225.	140.	53.	40-60.
Ti/Zr	80.	120.	59.	35.	78.	74.	84.	160.	65.	75.	130.	70-120.

**For basaltic-andesite

or, at worst, 15%. The variations observed are not, therefore, analytical accidents. Normative assemblages are dependent on iron oxidation. By setting $Fe_2O_3/FeO = 0.25$ I have not allowed for primary variations in the oxidation state of third period rocks. When measured, shoshonitic series Fe_2O_3 contents especially seem in excess of what I assume in normative calculations. It is difficult to assess the magnitude of this effect but it is not likely to be large. Alteration due to weathering and burial metamorphism can affect Na, K, Rb, Ba, Sr, Th, and U whole rock concentrations (section 4.2D) and many of my samples have volatile contents (LOI) in excess of that considered maximum for unaltered rocks by, for example, Hart and Nalwalk (1970). My only check against such random alteration effects is consistency within suites, discussed below. It would, however, be extremely fortuitous for these regional variations to be alteration effects alone.

Second, my samples must be representative volcanic rocks of the stratigraphic units from which they were collected. As discussed in section 1.5, I chose samples thought representative of outcrop areas, not rock types, discriminating only by criteria of apparent freshness. Again, however, consistency is the only safeguard. Thus, one is less sure how to interpret areas for which few analyses are available (e.g. Kadavu, and most Lau, Lomaiviti, Mamanucu, and Yasawa islands).

Third, there must be more polarity-significant geochemical variability between than within the rock units or islands compared. My analyses do not always define stratigraphic units with internally consistent geochemistry. This is most evident for K and Rb contents of some Mathuandrove Super-Group rocks from central Vanua Levu (section 5.2C), but is also apparent in analyses of some of the Lau Volcanics (section 6.2C). Whatever the age of analyzed samples from the Mamanucu and Yasawa islands, they are themselves variable and most are different in composition from potentially correlative volcanics in Viti Levu.

Present island arc volcanism is also variable in composition, even at the same or adjacent craters over short time spans. The variations documented in section 3.2 or by Dickinson and Hatherton or Gorshkov are summaries of gross trends to which there are many local exceptions. I regard these differences in Fiji only as exceptions to the regularity described above for several reasons. The majority of analyses from each unit define the trend I cite so that the rest are, numerically, exceptional. These exceptions are almost all acid andesites or

dacites, i.e. rocks in which these elements could easily be concentrated by fractionation processes of local significance only. Moreover, at least in Vanua Levu, they appear to be spatially restricted suggesting limitation to one volcano or volcanic event. Finally, the exceptions are not equally different in all elements. For example, Vanua Levu rocks with higher than average concentrations of K, Rb, and Ba still have less Sr, Th, and U and more Y than rocks similar in age and SiO_2 content on Viti Levu.

Fourth, my proposal loses support unless the stratigraphic correlations I have adopted are correct. Obviously it requires that the Mathuandrove Super-Group, Namosi Andesites, Lau Volcanics, and analyzed samples from the Kadavu Group be Late Miocene or at least equivalent in age. Such designation of the first two units is paleontologically sound; of the third unit it is contentious; and of the last group it is blind faith. Less obviously, the simplicity but not necessarily the validity of my proposal is marred if samples 860 and 862 are representatives of the Late Miocene Nandi Group, if Tholo plutonism is only 7 to 13 m.y. old in eastern Viti Levu, or if Tavua volcanism is as old as 7 to 10 m.y. (The latter two sets of ages were reported by Rodda et al.,[1967]. See the discussion in sections 4.2C, 4.5A, and 4.2A, respectively, for reasons why I think none of these is true.) The simplicity is apparently compromised in any case by the association of hornblende-andesites with third period basalts and shoshonites in northern Viti Levu, Beqa, Ovalau, Nairai, and Vanua Balavu (sections 4.4E and F, 6.4B, and 7.2).

Fifth, the regional variations described in chapter 3 and apparent in Table 8.1, and the differences in mafic compositions listed in Table 8.2 must be more than the effects of crystal fractionation or crustal contamination. The concentration of most polarity-significant elements increases during fractionation because they are excluded from crystallizing phases. I have tried, therefore, to avoid comparing the minor and trace element contents of rocks with very different major element compositions. There is no a priori reason to expect variable crustal contamination as all islands lie on the Lau-Colville Ridge and few differences in thickness or age have been demonstrated or are expected within it. Moreover, as will be discussed in section 10.8, high concentrations of large cations in Kadavu sample 390 cannot result from crystal fractionation of any known Fijian magma type or from contamination by any likely crustal material.

Finally, the Lau-Colville Ridge must have evolved as a single, not composite, physiographic feature and its islands must approximately retain their previous geographic relationship to one another. Otherwise the regional variations observed are accidental. Chase (1971) and Malahoff (preprint) propose differential movements between islands, but I see no bathymetric or geologic evidence for major structural dislocation within the ridge. Moreover, I regard the similarities in age of volcanism on different islands and the regularities of geochemical variations between the composition of these volcanic rocks of similar age too consistent to be accidental.

The principal geochemical data not explained by this proposal are the differences between the Lau Volcanics and rocks of the Mathuandrove Super-Group, especially their different 87Sr/86Sr ratios.

Running throughout my argumentation has been an assumption about what is normal and abnormal volcanism accompanying subduction. "Normal" in the sense of most common, means being basaltic-andesite or andesite in composition and/or having less than 1.2% TiO_2. As an empirical statement this is easily verifiable. Equally verifiable, however, are exceptions, usually in somewhat atypical island arcs like the West Japan Volcanic Belt, the Bismark Archipelago, or the New Hebrides, where direction of underthrusting is at an oblique angle to the trench or where inter-arc rifting contributes to volcanism. Because Fiji may easily have been an atypical area in pre-Pliocene time, use of the normality notion is questionable.

I have used it in two ways. First, I attribute pre-Pliocene volcanism in Fiji to "normal" island arc activity because rocks of this age are dominantly basaltic-andesite or andesite and have less than 1.2% TiO_2. In addition, Late Miocene rocks display the same regional variations in geochemistry observed in modern arc volcanism. Second, I consider the Pliocene to Recent shoshonites and basalts abnormal because they are dominantly mafic and some have $\sim$2% TiO_2. This assessment is less certain. However, because Fiji was once and is no longer a "normal" island arc, I choose to equate this change in magma composition with the change in plate relationships which has led to Fiji's present abnormality. Apparent equivalence in time between initiation of mafic volcanism in Fiji and age of the Lau Basin and parts of the Fiji Plateau supports this assumption.

8.4 Fiji's relationship to the New Hebrides and Tonga

The foregoing analysis hopefully demonstrates that the entire Lau-Colville Ridge was an island arc on the eastern border of the Australian plate beneath which the Pacific plate was thrust during the Late Miocene, and that this configuration changed during the Pliocene resulting in basaltic and shoshonitic volcanism in Fiji. Gorton's and my earlier correlation of Fiji's geologic history with that of Tonga and the New Hebrides (Gill and Gorton, Appendix 3) is probably premature and, at best, a proposal. Considerably more geological, geophysical, and oceanographic information is needed before confirmation or refutation is likely. It will be particularly important to establish the minimum age of massive intermediate volcanism on the Lau-Colville Ridge (i.e. to verify the age of the Mathuandrove Super-Group, Namosi Andesites, Lau Volcanics, and Kadavu high-K calc-alkaline rocks), to establish the maximum age of shoshonitic and basaltic volcanism and to assess variations in its composition with space and time, and to establish the maximum age and history of the Lau Basin and Fiji Plateau.

8.5 Implications for magma genesis in Fiji

There is a tendency but not necessity for the thesis to become circular at this point. If the proposals above are valid, then magma genesis during the first and second periods of Fijian volcanism occurred in an island arc environment, above down-thrust lithosphere. The geochemistry of pre-Pliocene volcanic rocks is, therefore, pertinent to the more general issues of island arc magma genesis and defines a continuum between the island arc tholeiitic series, as defined by Jakes and myself (1970), and the calc-alkaline and high-K calc-alkaline rock series. Pliocene volcanic rocks present a different continuum, between alkali and tholeiite basalt on one hand, and alkali basalt and shoshonite on another. Genetic models must explain the features of each and the gradation between these rock series.

Chapter 9. Boundary Conditions for Magma Genesis in Island Arcs with Special Reference to Fiji

9.1 Introduction

Magmas erupted in Fiji could have originated within the crust of the Lau-Colville Ridge, the mantle beneath it, or--at least during pre-Pliocene time--lithosphere thrust into that mantle (Figure 9.1). They may be (1) melts which rose directly from their place of origin to the surface with little modification en route, or (2) liquids which have (a) fractionated at their origin or any higher level, (b) been modified by reaction with rocks through which they passed during ascent and crystallization, or (c) both.

In an attempt to constrain these alternatives I have summarized in the following section salient points from data in chapter 3 and their sources, the reviews of island arc seismicity by Isacks and Molnar (1971) and Utsu (1971), and the preceding discussion of Fiji. In sections 9.3 to 9.5 I briefly evaluate the bulk composition, mineralogy, and melting behavior of the possible source regions listed above, and of the crystallization behavior of magmas within them. Together these suggest that intermediate magmas in island arc environments could form in many ways and that there is likely to be no unique and universally applicable solution to the "andesite problem". Thus, as in most aspects of geology, one must rely on empirical descriptions such as those in chapters 4 to 8 and degree of compatibility with other data to constrain choices. Chapters 10 to 12 are nascent efforts to see which options are most compatible with my observations. These evaluations proceed by trying to deduce from the geochemistry of a rock suite the mineral assemblage which, upon fractional crystallization or partial melting, would yield a suite with that composition. If successful, one could set limits on the depth, temperature, and water content of the magma source during fusion, or the depth, temperature, and water content of the magma during its last fractionation. Though possible in principle such deductions are difficult in practice. They require, among other things, that the rocks I analyzed were once entirely liquid and co-genetic within suites, which are assumptions I cannot prove and, frankly, doubt. More fundamentally, it is easier to disprove than prove by an empirical method, as no finite number of observations constitute proof. This is clearly illustrated in chapters 10 and 11.

9.2 General geochemical and tectonic constraints

Any explanations of island arc magma genesis must account for, among others, the twelve following observations.

(1) Most volcanoes of modern island arcs lie about 100 to 150 km above the Benioff zone. That distance ("h" for Dickinson and Hatherton) is never less than 80 km. Mantle between the volcanic front and trench is not seismically anomalous but that behind the volcanic front is characterized by low body-wave velocities and high attenuation (section 9.4). Mantle beneath active volcanoes is ambiguous, does not seem anomalous on a regional scale, but does seem to contain definite magma chambers (Fedotov, 1968).

(2) Most volcanic fronts maintain strict parallelism with their associated trench and with contours of well-located earthquake foci.

(3) The composition of island arc magmas is predominantly intermediate and changes regularly with distance from the trench as shown in Figures 3.1 to 3.3. Hatherton and Dickinson (1969) and Sugimura (1972) demonstrated statistically significant correlations between depth to Benioff zone seismicity and both K_2O contents and normative mineralogy in island arc volcanics. (The former are better correlated in intermediate rocks, the latter in basic.)

These three factors suggest an intimate relationship between underthrusting and volcanism but do not require magma genesis at or near the slab-mantle interface. The first point, when coupled with factors to be discussed in section 9.5, requires that if magmas originate within underthrust lithosphere and rise vertically, they must represent partial melts of eclogite. It may also indicate that anomalous mantle behind island arc volcanic fronts is <u>not</u> due to slab dehydration (see section 9.5).

(4) Major and minor element compositions may or may not change along strike as magmas rise through thicker crust. Gorshkov (1962, 1970) stressed that they do not in the Japan and Kurile-Kamchatka areas; Forbes <u>et al</u>. (1969) and Brothers (1970) show that they do change in the Aleutians-Alaska and Kermadec-New Zealand analogies, respectively. From data in Gorshkov (1970) and Kravchenko and Semenov (1970) there appear to be northward increases (i.e. corresponding to areas of thicker crust) of K, Th, and U contents in the Kuriles although this is apparently not thought significant by Gorshkov and not apparent in the Th, U, Rb, and Cs data of Leonova and Udaltzova (1970).

(5) Sr isotope ratios are too low for island arc magmas to have incorporated much low melting fraction from old crust with high Rb/Sr ratios (Gill and Compston, Appendix 3). They are not too low to exclude contamination from young island arc crust but this begs the question of how that crust acquired its isotopic character.

(6) Sr and Pb become less radiogenic with distance from the trench (or above Benioff zone seismicity) in the East Japan Volcanic Belt, the Late Miocene of Fiji, and possibly modern New Zealand. If this is a fundamental and consistent characteristic, it minimizes the possibility that lateral variations in island arc magma composition could be due to crustal contamination or mantle wall-rock reaction (see section 10.8). Data is insufficient, however, to establish its generality. Page (pers. comm., 1971) found no Sr isotope variations in New Britain volcanics nor did Church and Tilton (in prep.) in the Cascades, although the latter did find a minor increase in 206Pb/204Pb and 208Pb/204Pb ratios with distance from the continental margin. For the West Japan Volcanic Belt, Kurasawa (1970) found that Sr became more radiogenic with distance from the trench whereas Tatsumoto and Knight (1969) found that Pb became less radiogenic. As noted in chapter 3, both include analyses of high-Ti basalts in their studies.

(7) High-K calc-alkaline rocks are found in areas of modest crustal thickness (≷25 km) and youth (Tertiary): e.g. Kadavu and Ono-i-Lau (Fiji), Bougainville (Solomons), the Volcanic Islands (N. Marianas), and Bogoslof (Aleutians). However, most rocks of this type are confined to areas of older, thicker crust.

(8) Extensive ignimbrite fields usually have a distinctive geochemistry (Jakes and White, 1972), higher 87Sr/86Sr ratios than most island arc volcanics (e.g. Ewart and Stipp, 1968), and are limited to areas with thick (>30 km), pre-Tertiary crust.

These five factors suggest variable crustal influence on island arc volcanism. Its variability, however, suggests that the crust is not the primary source of the consistent geochemical variations observed.

(9) Sr isotope ratios in island arc volcanic rocks are too high to reflect partial melting of pristine ocean floor basalt (Gill and Compston, Appendix 3; Hedge and Lewis, 1971; Hart, 1971b).

(10) Temporal variations in island arc geochemistry are apparently accidental.

(11) Predominantly basaltic (<52% SiO_2) and shoshonitic volcanic provinces are confined to areas which have relatively thick crust and/or are associated with inter-arc rifting or complex tectonics.

(12) Even intermediate volcanism may not always be related in a simple way to underthrusting. North-dipping lithosphere is not likely to underlie volcanoes north of New Guinea (Morgan, 1966; Johnson, 1970). Unless lithosphere is being thrust south under New Guinea as well as north under New Britain, as proposed by Jakes and White (1969) and Johnson and Molnar (1972), this volcanism is unsupported by underthrusting. Complex plate relationships certainly exist in that area, but oppositely dipping underthrusting renders fortuitous the apparent volcanological continuity between New Britain and the north coast of New Guinea.

9.3 Melting and crystallization within the Lau-Colville Ridge (region I in Figures 9.1 and 9.2)

As discussed in chapters 2 and 4 to 8, the Fiji islands are pinnacles of the Lau-Colville Ridge which is at least 30 km thick at 16°S and 15 km thick at 31°S. Governmental prohibition of explosion seismology precludes knowledge of velocity profiles beneath the major islands but Shor et al. (1971) report 7.0 km/sec velocities for the lower 8 km of crust at 31°S, which could indicate rocks of either amphibolite or granulite facies. These rocks may have been island arc volcanics and volcaniclastics, crystal residues from low-pressure fractionation, Mesozoic crust once part of the Norfolk Ridge, or oceanic sediments underplated and/or accreted during subduction (section 4.2A). Pre-Eocene oceanic crust probably immediately overlies mantle at the base of the Ridge. Most volcanic xenoliths are gabbroic, could be cognate, and reveal little about underlying crust.

No heat flow measurements have been published for Fiji. Radioactive heat production calculated using the factors of Birch (1954) and my K, U, and Th data for basement volcanic and volcaniclastic rocks would be about 0.1 $\mu cal/cm^2$ sec for 30 km of such material. This is similar to Sass and Munroe's (1970) estimate for crustal contribution to heat flow in the Aleutian and Philippine islands. Unless significant amounts of highly fractionated crust drifted eastward to underpin modern Fiji, temperature gradients within the Lau-Colville Ridge would therefore have differed from 20°/km (taken from Clark and Ringwood, 1964) only when affected by descending cold lithosphere or ascending magma. Thus, unless magma was introduced

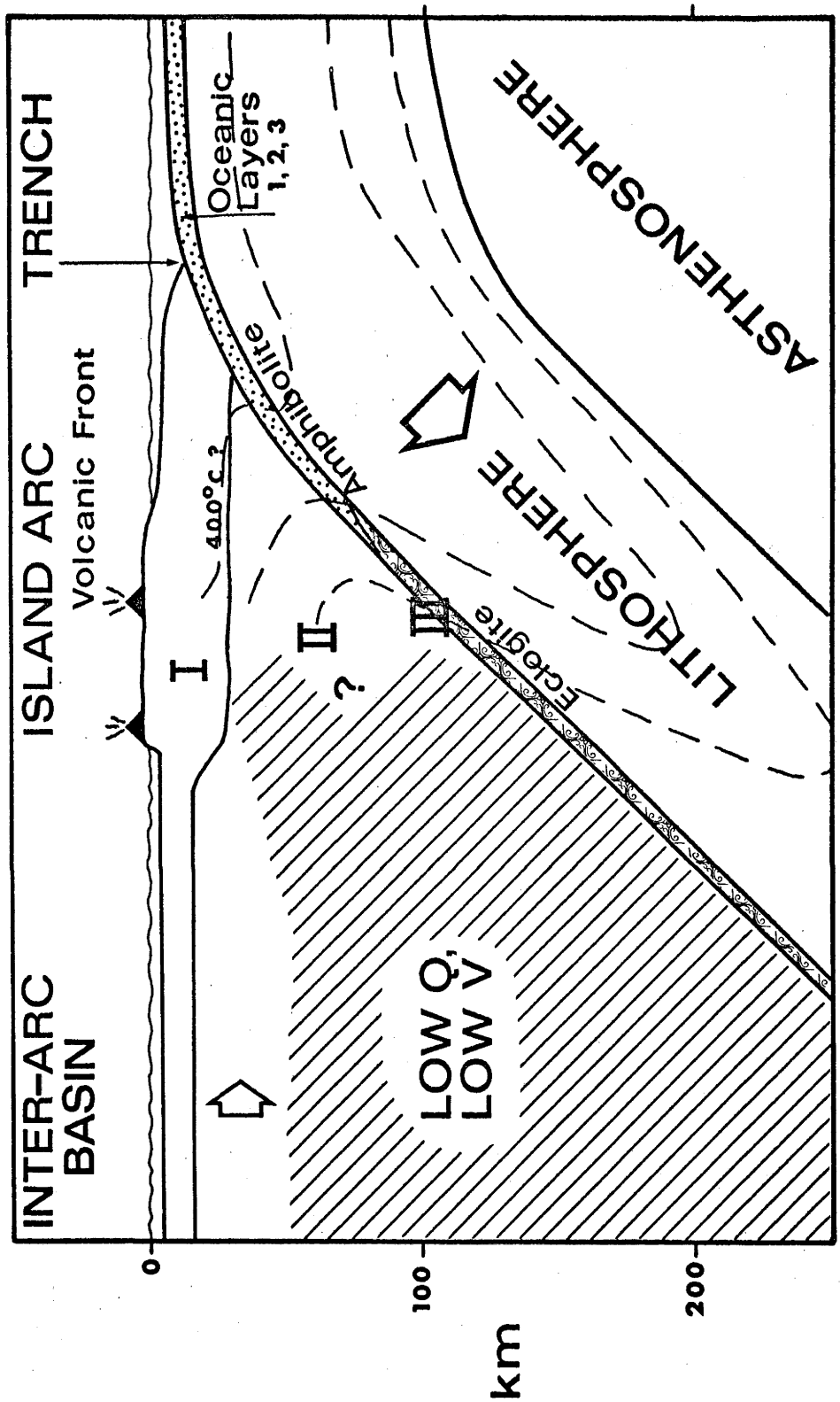

Figure 9.1 Tectonic structure of island arcs and possible magma source regions (I, II, and III). Isotherms and the spatial distribution of low-Q, low-V upper mantle are schematic. Dotted portion of oceanic crust indicates approximate region in which amphibolite is stable; patterned portion indicates similar region for eclogite. Arrow beneath inter-arc basin indicates inferred direction of dilation and arc migration.

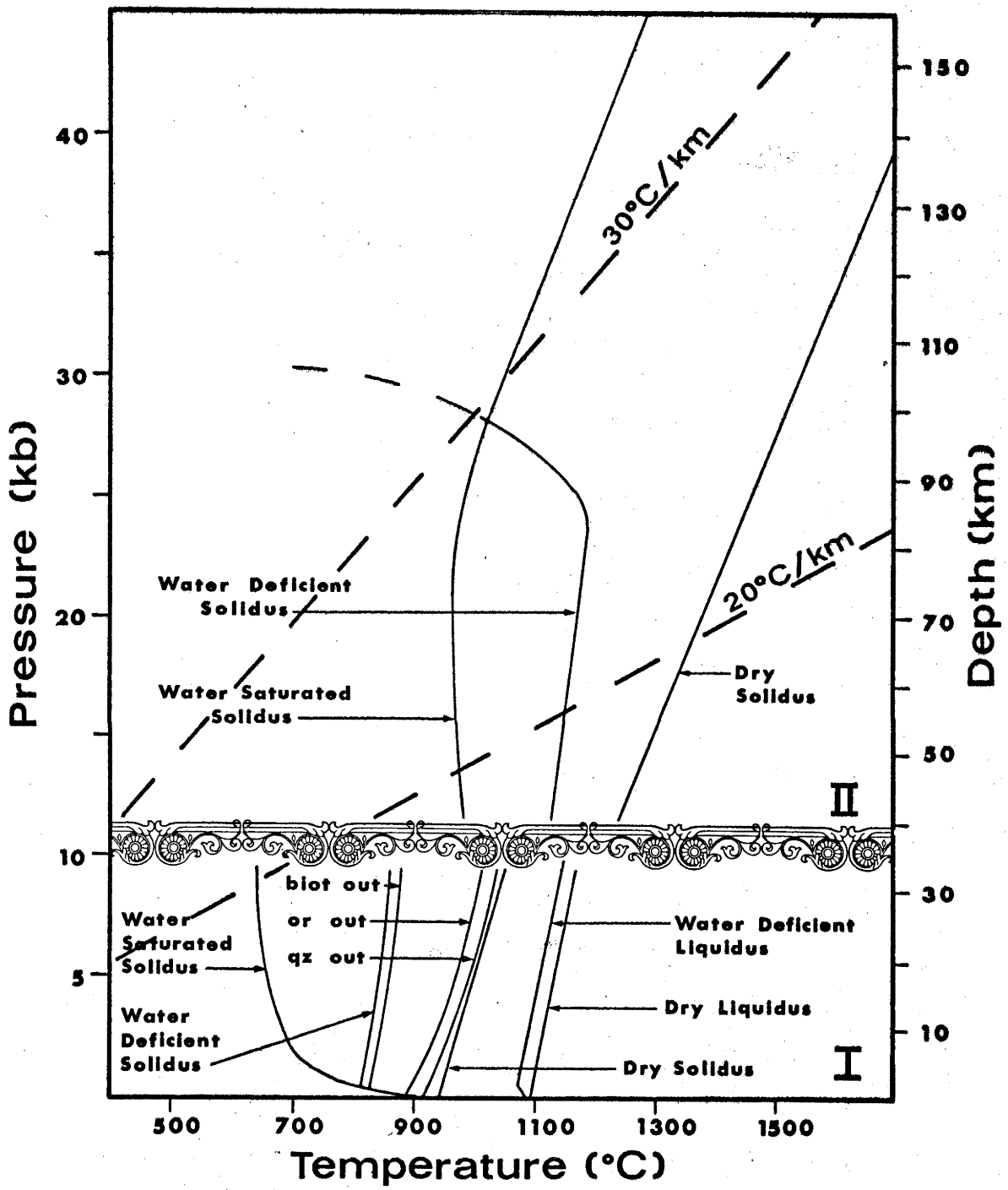

Figure 9.2 Generalized diagram for the melting of crust (region I) and mantle (region II) beneath island arcs. Region I: melting relations of a granodiorite composition from Robertson and Wyllie (1971) and Burnham (1967). "Mineral out" curves are for water-deficient systems. Note inflection in the water-deficient liquidus. Region II: melting relations of pyrolite composition from Green (1972). For both regions, "water-deficient" is used to designate systems in which the subsolidus assemblage is undersaturated with water, i.e. in which no water vapor phase is present.

from elsewhere, temperatures within the crust could not exceed even the water-saturated solidi of gabbros, granodiorites, or amphibolites and melting could not occur (see Figure 9.2). The source of most (but not necessarily all) Fijian magmas must lie below the crust.

Magmas may, however, crystallize, react with, and/or cause melting within this crust. Temperatures in the lower crust of island arcs are estimated to increase by 650 to 900° beneath the volcanic front (Uyeda and Horai, 1964; Oxburgh and Turcotte, 1971) but absolute values are difficult to assess. Local temperatures above the relevant solidus are therefore likely. Assuming volatile contents less than those of surficial rocks, a water-deficient subsolidus assemblage with or without a free vapor phase will occur. Melting at 8 kb (depths of ∿30 km) will begin between 700 and 900°C depending on bulk composition and volatile content (Burnham, 1967; Robertson and Wyllie, 1971, and references therein: see Figure 9.2). Although the width of this melting interval depends on the source composition (e.g. is greater for tonalitic than granitic sources) and its water content, the initial composition of the resulting magma is less so because it will co-exist with quartz, alkali feldspar, plagioclase, and ferromagnesian minerals until the quartz and alkali feldspar dissolve (Figure 9.2). The more water-saturated such melts are, the less vertical movement is possible without crystallization. Brown and Fyfe (1970) provide microprobe analyses of glass formed experimentally under these conditions from partial fusion of various crystal materials. They resemble analyses of Tholo Plutonic quartz-tonalites except that the latter contain more anorthite and ferromagnesian and less orthoclase components (Table 9.1). This might be expected if the crust being melted had no more K_2O than first period Viti Levu rocks. Acid Tholo plutons might, therefore, represent partial melts of Lau-Colville Ridge crust which were too water-saturated to reach the surface. Trace element problems with this interpretation are discussed in section 4.5C; implications for crustal evolution are mentioned in section 13.C.

The possibility of anatexis also means possible alteration of magmas passing through or crystallizing within the crust. If Viti Levu quartz-tonalites represent a low-melting fraction, then the geochemical effects of crustal contamination in Fiji can be estimated using an average Tholo quartz-tonalite composition (Table 9.1). A major contribution from such material would be more difficult to detect from

Table 9.1 Average Composition of Tholo quartz-tonalite and Experimentally Produced Anatectic Melts

	Average Tholo Quartz-tonalite	Average Anatectic Melt*
SiO_2	72.7	72.8
TiO_2	0.4	0.4
Al_2O_3	13.9	15.2
Fe_2O_3**	3.2	1.8
MnO	0.1	0.1
MgO	0.8	0.6
CaO	3.1	0.6
Na_2O	4.8	4.8
K_2O	0.9	3.7
P_2O_5	0.1	-

Rb	9.	V	32.
Sr	160.	Cu	5.
Ba	210.	Sn	2.
Pb	2.	Hf	4.
Th	2.0	Zr	125.
U	0.7	Na_2O/K_2O	2-20.
La	11.	K/Rb	900.
Yb	2.3	Rb/Sr	0.06
Y	35.	Ba/Rb	24.
Ni	-	Th/U	2.8
Co	5.	K/U	18,000.
Cr	6.	Zr/Hf	31.
Sc	9.	87Sr/86Sr	0.7043

* Brown and Fyfe (1970)

** Total Fe as Fe_2O_3

trace element or isotope data than is the case for older continental areas (Taylor, 1969; Gill and Compston, Appendix 3), but is unlikely to explain the spatial variations observed in the composition of Fijian volcanic rocks.

The range of rock types within Fijian suites may represent fractionation within crustal magma chambers at pressures <8 kb. There is seismic evidence for chambers at these depths in Alaska (e.g. Matumoto, 1971), the Kuriles, and Kamchatka (e.g. Fedotov, 1968). If Fijian magmas fractionated as well as crystallized under these

conditions, removal of minerals described in preceding chapters as phenocrysts in each suite is the most likely mechanism.

9.4 Melting and crystallization within the mantle beneath the Lau-Colville Ridge (region II in Figures 9.1 and 9.2)

The composition, mineralogy, and melting behavior of and crystallization behavior of magmas within the upper mantle have been extensively studied (see reviews by Ringwood, 1966, 1969; Green and Ringwood, 1970; Green, 1971; Anderson et al., 1971; Wyllie, 1971a, chapters 6 and 8; and others). These authors summarize estimates of mantle composition and its subsolidus mineralogy and melting relationships under various conditions. Mantle temperatures are unlikely to exceed the solidus of dry peridotite.

Mantle behind and possibly beneath island arcs differs in some respects from that elsewhere and has also been extensively studied (see review by Utsu, 1971). It is aseismic, bounded above by seismicity associated with magma chambers at 20 to 40 km depth and below by the Benioff zone, and characterized by high attenuation (Q = 50 to 100) and low velocities (~7½ km/sec) for P waves. Mantle beneath the Lau-Colville Ridge (i.e. beneath seismic stations on Viti Levu, Vanua Levu, and Lakeba [Lau] islands) does not share these characteristics today and, instead, transmits P and S waves with normal velocities and attenuation (Mitronovas and Isacks, 1971; Barazangi and Isacks, 1971). This will be significant when discussing the origin of Pliocene to Recent volcanism (chapter 12) but, by uniformitarian inference, upper mantle beneath Fiji in pre-Pliocene time probably had the characteristics described above for that beneath modern island arcs.

The areal extent of anomalous sub-arc mantle is unclear but in the most-studied areas (Tonga, Japan, Kuriles, Kamchatka) it begins at or behind the region of active subaerial volcanism. Utsu (1971, Figure 16) and Barazangi and Isacks (1971, Figure 17) conclude that high-Q, high-V mantle occurs above the Benioff zone between the volcanic front and trench (i.e. above seismicity shallower than 80 to 100 km), and that low-Q, low-V mantle occurs immediately above the Benioff zone at depths between 100 and 200-300 km (without an intervening gap of normal mantle above the slab as supposed by Sugimura, 1972). Figure 9.1 illustrates this relationship.

Low-Q, low-V upper mantle is also encountered in the low-velocity zone which separates "lithosphere" from "asthenosphere" and is probably

the source region for most oceanic volcanism. There these properties are thought to reflect the presence of a few per cent of silicate liquid (Lambert and Wyllie, 1970; Ringwood, 1969; Anderson et al., 1971) formed due to the breakdown of hydrous minerals (Green, 1971; Wyllie, 1971b). Interstitial liquids and therefore incipient melting are also inferred to occur between the Benioff zone and depths as shallow as 30 to 40 km behind and possibly beneath island arcs (Aki, 1968; Karig, 1971b; Barazangi and Isacks, 1971; Farberov and Gorelchik, 1971).

There is a dispute as to whether or not these liquids beneath island arc environments are intermediate in composition. Kushiro et al. (1968), Kushiro (1970), and Yoder (1969) discuss experimental evidence that they can be for water-saturated conditions as pressures up to 25 kb, and propose that this also applies to water-undersaturated conditions. They infer that the intermediate magmas of island arcs can be primary partial melts of hydrous mantle at depths of up to 60 to 80 km (also see O'Hara, 1968). Green (1970) disagrees. He notes that if, as claimed by Kushiro and Yoder, enstatite melts incongruently at high pressure in the presence of water to yield forsterite + quartz-normative liquid, then that liquid must co-exist with olivine at high pressure (i.e. have olivine as a liquidus or near-liquidus phase). He then describes unsuccessful experimental attempts to find such olivine in basalts with 0-2.5% normative quartz at 22.5 kb even in the presence of a free water phase. Nicholls (pers. comm., 1972) recently determined the upper pressure limit of olivine as a liquidus phase for basalt exactly saturated with silica to be <5 kb when dry and 17 kb in the presence of excess water. It appears possible, therefore, that silica-saturated or even quartz-normative basalt may indeed be a primary melt from peridotite containing high (>3%) water contents at depths less than about 40 km, but it is very unlikely that magmas with >>52% SiO_2 can originate in this way.

Direct mantle derivation of most intermediate rocks in island arcs seems precluded by other aspects of their composition as well. Magmas in equilibrium with mantle olivine + clinopyroxene will have $Mg/Mg+Fe^{+2}$ ratios >0.65 (Green, 1970; Irving, 1972) and Ni, Co, and Cr contents greater than 100, 40, and 150 ppm respectively (Taylor, 1969). Most intermediate island arc rocks have $Mg/Mg+Fe^{+2}$ ratios between 0.5 and 0.6 and less than half the appropriate Ni, Co, and Cr contents.

Thus mantle between modern island arcs and their associated Benioff zone may contain a few per cent basalt but not intermediate magma (unless high pressure fractionation occurs). The existence of this magma requires the presence of water in sub-arc mantle (Figure 9.2). Composition of the basalt depends on the depth, temperature, and water content at the site of melting and on the degree of melting (which also depends on temperature and water content). D. Green (1971, 1972) and Wyllie (1971a) provide a comprehensive account of possibilities.

Similarly, fractionation of these primary basalt magmas may occur within sub-arc mantle. Indeed, the more water-rich a magma, the more it must crystallize during ascent. Liquidus phases likewise vary with depth, temperature, and water content (Green, and Wyllie, ibid.).

Finally, magma originating within or ascending through sub-arc mantle may change composition due to "wall-rock reaction" (Green and Ringwood, 1967), i.e. due to the breakdown of hydrous minerals to anhydrous products plus water and "incompatible elements" in the presence of silicate melt. This will occur either if water-undersaturated magma (in which $P_{H2O}<P_{total}$) comes in contact with water-saturated peridotite (in which $P_{H2O} = P_{total}$) or if magma influx causes temperatures in adjacent wall rocks to rise above the solidus fixed by their water content. The latter process will cause magma temperatures to decrease resulting in lower solubility of water and, perhaps, crystallization.

Green and Ringwood appealed to this mechanism to account for the concentrations in basalts of minor and trace elements such as K, Ti, P, Rb, Ba, Cs, Sr, Pb, Th, U, Zr, Hf, Nb, and the light REE. Because of their ionic radii and charge, these elements do not occur in lattice sites of major minerals of peridotite (viz. are "incompatible") and probably exist mostly in anhydrous accessory minerals, along grain boundaries or, where stable, in amphibole or mica. Those not occurring in refractory accessory phases will therefore be enriched in magmas during wall-rock reaction as defined above.

This proposal is difficult to test quantitatively (Gast, 1968). If understood as equivalent to breakdown of hydrous minerals it amounts to addition of something between water containing those elements more abundant in amphibole or mica than clinopyroxene, and the minimum melting fraction of peridotite. This probably means approximately nephelinitic compositions (Bultitude and Green, 1971) rich in the elements listed above, as they are all more abundant in amphibole and phlogopite than clinopyroxene in high-pressure assemblages (see Varne

and Graham, 1971; and Griffin and Murthy, 1969, for example). Under such conditions equilibrium may not occur and the behavior of minor minerals such as apatite, zircon, Fe-Ti oxides, or sulfides will appreciably affect wall-rock contribution to adjacent magma.

9.5 Melting and dehydration behavior of underthrust lithosphere (region III in Figures 9.1 and 9.3)

Isacks et al. (1968) and several subsequent authors have demonstrated unequivocally that Benioff zone seismicity beneath island arcs occurs within an area of high body wave velocities and low attenuations, and that this area is an inclined slab (viz. "underthrust lithosphere") extending from the surface to depths of about 650 km. This slab acts as an effective wave guide and can be demonstrated to be continuous with lithosphere lying on the opposite side of the intervening ocean trench from island arcs, as shown in Figure 9.1. It seems therefore that lithosphere (i.e. oceanic crust plus mantle) about 100 km thick underthrusts modern island arcs, and that Benioff zone seismicity is the result of stress release within a relatively narrow (5 to 25 km wide) band near or at the top of that lithosphere. This rigid slab is a lowermost boundary for possible source regions of island arc magmas, is itself a potential source, and its behavior during underthrusting will influence magma genesis in overlying mantle. This behavior, in turn, depends on the composition, temperature, and water content of the slab.

Several authors have speculated on its thermal structure. McKenzie (1970) and Hasebe et al. (1970) calculated the temperature distribution (in dimensionless units) for a geometry of underthrusting similar to that shown in Figure 9.1 where I also include their schematic results. Toksöz et al. (1971) and Oxburgh and Turcotte (1970) tried to specify absolute temperatures. This requires setting temperatures at various depths equal to those of a chosen basalt solidus and quickly provides circular arguments about the possibility of partial melting within the slab. Note several aspects of these authors' calculations. (1) Temperatures within the slab are lower than within surrounding mantle. (2) Temperatures along the fault surface between the slab and overlying mantle may or may not be quite high depending on assumptions about shear-strain heating (compare figures of Minear and Toksöz, 1970, or Figure 14 of Oxburgh and Turcotte, 1970, with those of Toksöz et al., 1971, and see the criticisms by Sass, 1971). Inevitably, however, the upper part of the slab is its hottest portion due to conductive heating by the overlying mantle. (3) Isotherms are inverted within the upper

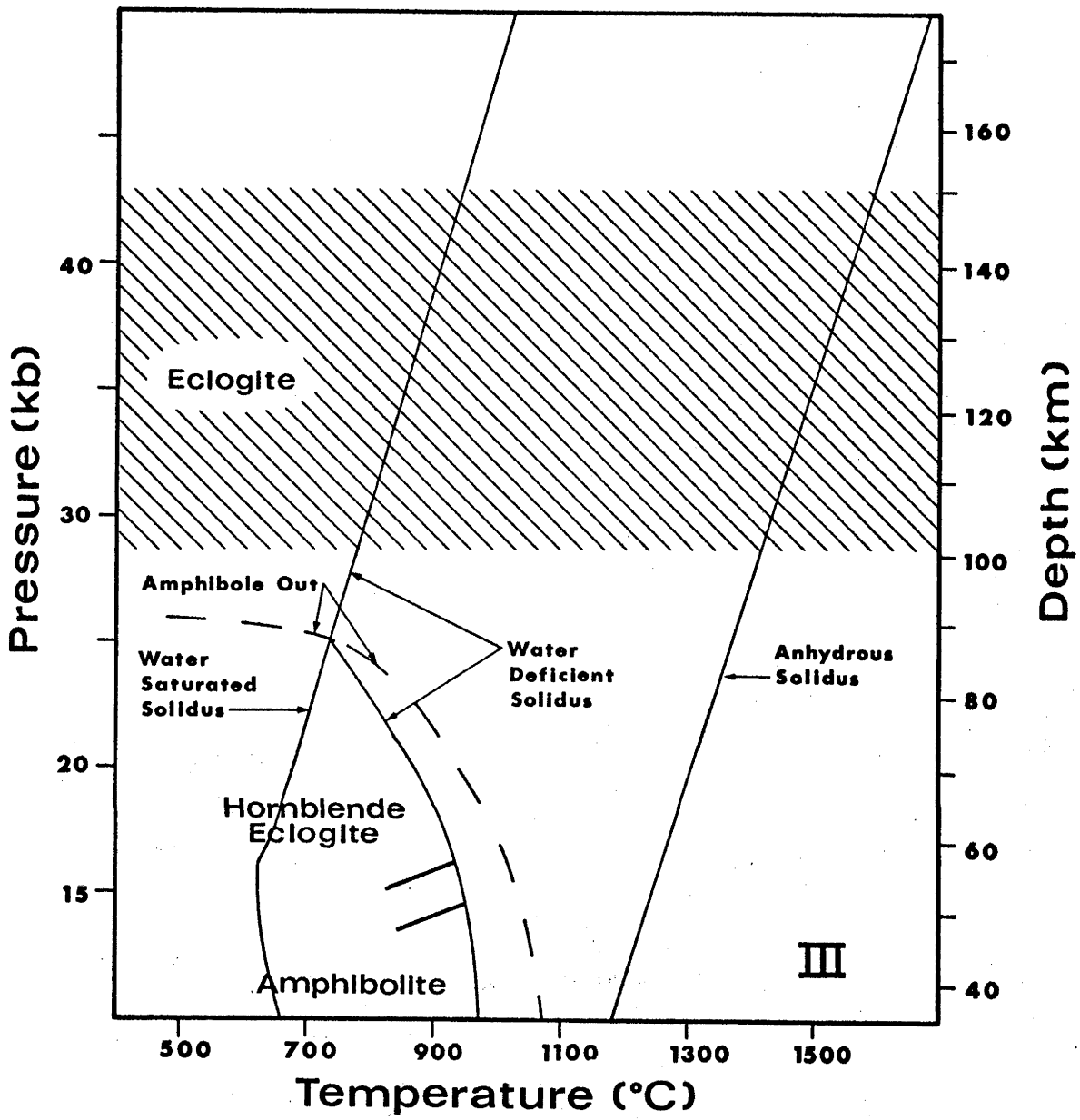

Figure 9.3 Generalized diagram for melting of the ocean floor basalt component of underthrust lithosphere (region III). Depths beneath most island arc volcanoes are indicated by diagonal lines. Phase relationships from Hill and Boettcher (1970), Holloway and Burnham (1972), Lambert and Wyllie (1970), and Wyllie (1971b). The "water deficient solidus" is for conditions of $p_{H2O}^{hb} = p_{H2O}^{melt}$. The amphibolite to hornblende-eclogite transformation is shown with an intervening transitional zone in which jadeite and garnet appear at the expense of plagioclase. Magma forming at the water deficient solidus is acidic; magma forming at the anhydrous solidus is intermediate (Green and Ringwood, 1968; T. Green, 1972).

slab and immediately overlying mantle so that rising diapirs or magmas initially encounter hotter, not cooler environments.

Whether melting occurs under these P-T conditions depends on the bulk composition and water content of the upper slab. Because the slab is laterally continuous with oceanic crust, its upper portion is probably made of similar material. Seismic refraction studies of Pacific ocean crust are summarized by Shor et al. (1970) and interpreted by many (e.g. Christensen, 1970) in the light of JOIDES drilling to indicate variable amounts of unconsolidated sediments (layer 1) underlain by basalt (layer 2) and amphibolites, gabbros, and/or serpentinites (layer 3; including the 6.9 to 7.6 km/sec layer identified by Maynard, 1970). The latter two layers are thought to have the chemical composition of ocean floor basalt. No one knows the fate of layer 1 during subduction. It is inferred to be scraped off, metamorphosed, and added to island arcs (Dickinson, 1971c; Ernst, 1971; Oxburgh and Turcotte, 1971) but it and/or its water may accompany layers 2 and 3 into the mantle. Mantle beneath oceanic crust is probably refractory, will have higher melting temperatures than overlying crust even if not, and is cooler during subduction. Therefore, melting will be localized in layers 2 and 3 and will occur if the solidus for basalt compositions is reached.

Figure 9.3 illustrates the following remarks. As outlined by Ernst (1971) and others, rocks of layers 2 and 3 will increase in metamorphic grade through several dehydration reactions during their subduction, leading to amphibolite, hornblende-eclogite or garnet-amphibolite, and eclogite facies assemblages. The drier or less silica-saturated the bulk composition and the lower its Ab/An+Ab and $Mg/Mg+Fe^{+2}$ ratios, the shallower the depth of eclogite formation (Green and Ringwood, 1972). Magmas can form at depths as shallow as 30 to 50 km if rocks are saturated with water, and would be in equilibrium with amphibole, clinopyroxene, feldspar, and possibly olivine. If instead, $P_{H2O}<P_{total}$, melting may not occur until pressures exceed those in which amphibole is stable relative to garnet + pyroxene + water ($\sim$25 kb), and then only if temperatures are above $\sim 700^{o}C$. Magma so formed will probably have high water contents, be initially quartz-rich (i.e. rhyodacitic: Green and Ringwood, 1968), become more mafic as degree of melting increases, and may involve non-equilibrium processes (Gill and Compston, Appendix 3). This will be discussed further in section 11.5B.

Between depths of about 70 to 90 km, amphibole will break down and garnet will occur in rocks of layers 2 and 3. If temperatures are below 700°C, melting will not take place and eclogite facies assemblages will stabilize. The subsolidus eclogite assemblage depends on bulk composition, Fe oxidation, and the grossular content of garnet (which increases with pressure: Green and Ringwood, 1968). As Fe becomes reduced, more quartz and less kyanite can form; as the grossular content of garnet increases, the diopside/garnet ratio decreases and more quartz can form. Dependence on bulk composition is illustrated in Table 10.1 for various estimates of the composition of layers 2 and 3. Magmas forming at depths greater than 100 km will be partial melts of eclogite, water-poor and therefore less subject to crystallization during ascent, intermediate in composition, and in equilibrium with garnet and pyroxene (Green and Ringwood, 1968; T. Green, 1972). Eclogite melting is discussed further in chapter 10.

As discussed above, seismic properties of mantle beneath some island arcs imply that it is partially molten and this, in turn, requires that it contain water. Dehydration reactions in the slab can provide copious amounts of water. If layers 2 and 3 are together 7 km thick (Shor et al., 1970), contain about 1% water (equivalent to maximum amphibole content in basalt: Hill and Boettcher, 1970), and are underthrust at 5 cm/yr, then each square km of lithosphere will, before reaching a depth of 80 km, exhale enough water to saturate 50 km^3 of overlying mantle (assuming saturation = 0.3% water: D. Green, 1972). If temperatures in the hotter overlying mantle are ∿1000°C, melting will occur (Figure 9.2). Thus, dehydration reactions alone could cause magma generation in the upper mantle. If, for example, mantle temperatures were ∿1150°C, 20% melting could occur in pyrolite with 0.3% water (D. Green, 1972, Figure 3) which could mean 500 km^3 of magma per 10^6 years per km^2 of underthrust lithosphere. (By comparison, the total rate of eruption in the Kurile arc is 0.65 km^3 per 10^6 years per km^2 of underthrust lithosphere: data from Markhinin, 1968, assuming underthrusting at 8 cm/yr at 45° along a trench 1100 km in length.) Although such rates of magma production are unlikely for several reasons, they illustrate the effect.

As noted in sections 9.2 and 9.4, high-V, high-Q mantle is encountered between island arc volcanic fronts and trenches (i.e. above Benioff zone seismicity shallower than 80 to 100 km). Mantle in this

region cannot therefore contain small amounts of silicate magma which requires that the mantle be cooler than ~1000°C, contain <0.3% water, or both. However, slab dehydration should be essentially complete before it reaches 90 km depths (Figure 9.3). This may imply that processes causing large scale partial melting of the upper mantle are unrelated to dehydration of the slab, are more relevant to formation of inter-arc basins than island arc volcanism (Karig, 1971b), and reflect shear-strain heat sources. Otherwise, mantle which overlies underthrust lithosphere must move away from the trench as was suggested by McKenzie (1969).

The preceding sections demonstrate that melting could occur within either the downgoing slab or mantle overlying it, that "wall-rock reaction" can modify magma during its ascent through the mantle, and that secondary anatexis and/or "contamination" can occur within the lower crust. Partial melts of lithosphere will be intermediate to rhyodacitic in composition; water-bearing partial melts of the sub-arc mantle must crystallize and fractionate during ascent; and crustal anatexis can add rhyodacitic liquid to magmas ascending through or crystallizing within the crust. Results of many processes therefore converge on intermediate magma compositions. The following chapters discuss a few of them in some detail.

Chapter 10. Genesis of Fijian Calc-alkaline and High-K Calc-alkaline Rock Series

10.1 Introduction

The concept "calc-alkaline rock series", as normally used, has never been rigorously defined. Peacock (1931) coined the term but the criteria by which he distinguished such rocks from alkalic ones do not separate the calc-alkaline from what Kennedy two years later called the tholeiitic rock series. In both, CaO can equal $Na_2O + K_2O$ at 56 to 61 (or higher) per cent SiO_2, and fractionation can yield quartz-normative liquids having a reaction relationship with olivine. After discovery in 1939 of iron enrichment in the Skaergaard, the adjective "calc-alkaline" was left, by default, denoting all magmas which, upon differentiation, become enriched in normative quartz but less enriched in iron than do Skaergaard-like liquids which conform to the "Fenner trend". Considerable confusion has resulted from subsequent use of this adjective to describe the abundant but diverse volcanic and plutonic rocks of orogenic belts which share these minimal characteristics, and the need for more discriminating semantic tools has been evident for some time. Kuno's distinction between pigeonitic and hypersthenic rock series, Taylor's between various andesite types, and that proposed by Jakes and myself (1970) are simply attempts to isolate consistently different features which will separate rocks of different parentage.

The Namosi Andesites (section 4.3) are calc-alkaline in the sense that they are predominately high alumina (average = 17.5%) andesites with no iron enrichment (Figure 4.3), moderate K_2O (average = 1.2%), and trace element contents (Table 8.1) similar to those thought average for andesite by Taylor (1969) and for calc-alkaline rocks by Jakes and myself (1970). T. Green (1972) chose Namosi Andesite sample 66 (Table 4.3) for experimental work relevant to the general problem of calc-alkaline rock genesis. He concluded that the Namosi suite could best be explained by eclogite-controlled fractional melting or crystallization at depths >80 km, and that high-K calc-alkaline rocks (e.g. Kadavu sample 390, Table 7.2 and section 7.3) formed similarly at greater depths (>150 km).

10.2 Eclogite melting and Namosi Andesites: major elements

T. Green (1972, Figure 1) found garnet and omphacite as near-liquidus phases of sample 66 at pressures >25 kb under anhydrous conditions. The order of appearance of garnet and pyroxene below the

liquidus is reversed relative to the andesite composition studied by Green and Ringwood (1968, Figure 4) but the position of liquidus, solidus, and the stability field of quartz is similar. From this Green argues that Namosi andesites and basaltic andesites represent partial melts of dry uppermost lithosphere at ∿100 km, in equilibrium with garnet and pyroxene (i.e. originate through partial melting of eclogite). He also determined the sub-liquidus mineralogy for sample 66 plus two, five, and ten per cent water and found (1) a notable depression of the liquidus with addition of water and (2) that the stability field of garnet plus pyroxene expands at the expense of quartz at high pressure. He therefore suggested that acid differentiates within the Namosi Andesite suite (e.g. dacite 874) could result either from hydrous fractionation at pressures >25 kb of an earlier derived andesite, or from initial hydrous melting with an eclogite residuum at depths of ∿80 km (the low-temperature trough described by Green and Ringwood [1968, Figure 8] moves to more silica-rich compositions during hydrous melting).

If Namosi andesites were unfractionated partial melts of lithosphere, then some combination of their composition plus those of phases with which they are saturated at or near the liquidus at high pressure should equal the composition of their source material within the lithosphere. (This assumes that no minor minerals are refractory and that garnet and clinopyroxene do not melt incongruently to form another mineral plus liquid. $Omphacite_{ss}$ melting yields pyroxene with a different composition plus silica-undersaturated liquid [Switzer and Melson, 1969]; pyrope melts congruently at high pressure [Boyd and England, 1959] but $garnet_{ss}$ yields another garnet composition plus liquid upon melting. No other minerals have been observed at near-liquidus temperatures in high pressure experiments on intermediate rock compositions. These assumptions are re-considered in section 10.5.) To quantitatively evaluate this proposal I adapted the computer program described by Wright and Doherty (1970) which uses a least squares method to solve a set of over-determined linear equations. Petrochemical significance rather than analytical precision was used to weight oxides.

As discussed in section 9.4, uppermost lithosphere (oceanic layers 2 and 3) is the source material most likely to occur in abundance in eclogite facies beneath island arcs and will probably have the composition of ocean floor basalt plus whatever is left of oceanic

Table 10.1. Estimates of average compositions of ocean floor basalts and sediment.

	(1) Basalt	(2) Basalt	(3) Basalt	(4) Basalt	(5) Sediment
SiO_2	49.21(0.7)*	49.34(3.1)	49.61(0.7)	50.6	60.45
TiO_2	1.39(0.3)	1.49(0.2)	1.43(0.3)	1.6	0.86
Al_2O_3	15.81(1.5)	17.04(1.7)	16.01(0.8)	15.1	18.28
Fe_2O_3	2.21(0.7)	1.99(1.2)	2.00**	4.3	-
FeO	7.19(1.2)	6.82(1.2)	9.69**	6.0	8.48(TOT)
MnO	0.16(.03)	0.17(.03)	0.18(.04)	0.18	2.20
MgO	8.53(2.0)	7.19(1.4)	7.84(0.9)	6.4	3.74
CaO	11.14(0.8)	11.72(0.9)	11.32(0.6)	10.9	0.79
Na_2O	2.71(0.2)	2.73(0.4)	2.76(0.2)	2.7	1.43
K_2O	0.26(.17)	0.16(.38)	0.22(.12)	0.47	2.97
P_2O_5	0.15(.04)	0.16(.22)	0.14(.07)	0.21	0.79
ECLOGITE NORMS* **					
Ap	0.3	0.3	0.3	0.3	-
Rut	1.4	1.5	1.4	1.6	-
Or	3.3	2.2	2.2	5.6	-
Ac	6.5	5.5	5.5	12.9	-
Jd	12.1 } 41.7	12.9 } 50.0	12.9 } 39.3	5.6 } 53.5	
Di	23.1	31.6	20.9	35.0	
Ga	43.9	29.1	51.5	20.2	-
Ky	3.4	10.8	1.1	13.8	-
Qz	5.8	5.4	6.1	7.7	-
Garnet Composition	$Alm_{25}Py_{43}Gr_{31}$	$Alm_{27}Py_{42}Gr_{31}$	$Alm_{32}Py_{37}Gr_{31}$	$Alm_{27}Py_{42}Gr_{31}$	
Omphacite Composition	$Jd_{29}Ac_{16}Di_{55}$	$Jd_{26}Ac_{11}Di_{63}$	$Jd_{33}Ac_{14}Di_{53}$	$Jd_{10}Ac_{24}Di_{65}$	
TRACE ELEMENTS					
	(6)	(2)	(7)		(5)
Rb	7(3)	-	1.8	-	110.
Sr	123(46)	130(25)	110.	-	180.
Ba	12(8)	14(7)	13.	-	2300.
Pb	-	-	0.5	-	80.
Th	-	-	≤.15	-	7.
U	-	-	.08-.30	-	1.3
La	-	-	2.7	-	115.
Yb	-	5(1.5)	2.9	-	15.
Y	43(12)	43(10)	30.	-	90.
Ni	123(56)	97(19)	120.	-	225.
Co	-	32(3)	32.	-	74.
Cr	296(80)	297(73)	300.	-	90.
Sc	-	61(19)	60.	-	19.
V	289(73)	292(57)	290.	-	120.
Hf	-	-	2.5	-	4.1
Zr	100(42)	95(35)	100.	-	150.

Data Sources:

(1) Melson et al. (1968)
(2) Engel et al. (1965)
(3) Cann (1971)
(4) Hekinian (1971)
(5) Wedepohl (1969)
(6) Melson & Thompson (1971)
(7) Adapted for calculations summarized in Figures 10.1 and 10.2. Data from (2), (6), and other sources. Rb was chosen so that K/Rb = 1,000 when K_2O = 0.22%.

Notes:

*Numbers in parentheses are one standard deviation.

**Assuming 2% Fe_2O_3

***Assuming that garnets are 31% grossular (as found for an ocean floor basalt composition at 36 kb by Green, 1967, Table 7.9), that magnetite is unstable relative to acmite (Banno and Green, 1968), and that Fe/Mg ratios are the same in diopside and garnet. Orthoclase rather than biotite (=Or + 3(Mg,Fe)O + H_2O) is calculated.

Table 10.2. Mineral compositions used in calculations.*

	1. 36kb CPX	2. 36kb GAR	3. 27kb CPX	4. 27kb GAR	5. HB	6. HB	7. CPX	8. MAG	9. MAG
SiO_2	54.3	40.8	51.0	40.6	44.5	44.4	50.9		
TiO_2	.35	1.2	.65	1.3	2.0	1.9	.41	7.2	7.2
Al_2O_3	12.6	21.8	12.4	22.0	9.8	9.7	2.5	.20	1.2
Fe_2O_3					5.5	5.6	2.4		
FeO	4.6**	17.1**	7.8**	16.4**	7.9	7.8	7.2	83.3	83.7
MnO		.49	.18	.62	.06	.07	.50	.25	1.0
MgO	7.9	9.0	10.6	10.6	13.6	14.0	14.3	1.8	1.7
CaO	14.9	11.5	15.9	10.0	11.6	11.1	19.9		
Na_2O	4.6		2.3		3.8	3.8	.60		
K_2O	.05		.07		.19	.18	.08		
TOTAL	99.4	101.9	100.8	101.7	100.6***	100.7***	98.7	92.8	94.8
Rb					1.4	1.3			
Ba						42.	52.		
Sr					135.	128.	44.		
Ce					27.	27.	11.		
Pr					8.3	10.			
Nd					26.	29.	5.7		
Y					56.	53.	31.		
Ni					22.	28.	26.		
Co					61.	66.	35.		
Cr					48.	62.	369.		
V					435.	425.	154.		
Cu					30.	29.	208.		

*Data in columns 1 to 4 are microprobe analyses by T. Green; in columns 5 to 7 are XRF analyses of Mau Quarry andesite phenocrysts by P. Jakeš (1970); and in columns 8 and 9 are microprobe analyses also by Jakeš.

**Total Fe as FeO.

***Sum includes H_2O and F.

layer 1. Several authors (Engel et al., 1965; Melson et al., 1968; Cann, 1971) give average compositions for fresh ocean floor basalt which are compared in Table 10.1 together with their eclogite norms. Gill and Compston (Appendix 3) observed that if such fresh basalts were isotopically equilibrated, they could not upon partial melting provide Sr with the isotopic composition of that found in island arc volcanic rocks. However, altered, weathered, or burially metamorphosed ocean floor basalts could. Hekinian (1971) gives an average of seven altered Pacific floor basalts (i.e. the sort omitted from the other averages with which it is compared in Table 10.1). As a limiting case, the upper seven km of downgoing slab could have a composition somewhere amid the range of analyses in Table 10.1. (The effect of sediment is considered in section 10.4.)

Of Green's experiments, only the one conducted at 27 kb and 1205°C with 5% water yielded both garnet and clinopyroxene at liquidus temperatures (T. Green, 1972, Figure 3). The compositions of these minerals (microprobe analyses: T. Green, pers. comm., 1971) are given in Table 10.2 and used as residual phases in computations. In all other runs these phases did not occur together until ∿25°C or more below the liquidus so that liquids with which they co-existed did not have the composition of the starting material, rendering the exercise less valid.

Results of the calculations are given in Table 10.3. The closest approximation was to Melson et al.'s average ocean floor basalt and was obtained by combining 29% liquid having the composition of sample 66 together with 44% residual clinopyroxene and 27% residual garnet. Similar amounts of liquid and residual phases but less success is encountered when trying to match other estimates for the composition of oceanic layers 2 and 3. Also in Table 10.3 are results of similar calculations using liquids with the composition of basaltic-andesite 876 and dacite 874. Residual garnet and clinopyroxene compositions change during melting so that these combinations are inherently less likely to yield reasonable results. Nevertheless, equally close approximations of ocean floor basalt compositions occur. The sensitivity of this computational method is discussed by Wright and Doherty (1970). In my experience, the calculated amounts of liquid and residual phases can vary by less than five relative per cent without the approximation becoming significantly worse.

Table 10.3. Calculated source compositions and amounts of residual phases involved when deriving Namosi Andesites from ocean floor basalts.*

	1.	ba	a	d	2.	ba	a	d	3.	ba	a	d	3.	ba	a	d
SiO_2	49.9	50.4	60.5	50.6	50.0	50.8	51.0	51.0	49.1	49.4	49.6	49.6	51.6	51.8	52.1	52.2
TiO_2	1.4	0.8	0.8	0.8	1.5	0.8	0.8	0.8	1.4	0.9	0.9	0.9	1.6	0.8	0.8	0.8
Al_2O_3	16.0	16.4	16.2	16.2	17.3	16.8	16.6	16.5	15.9	17.0	16.9	16.9	15.4	16.6	16.3	16.3
FeO**	9.3	9.3	9.5	9.5	8.8	9.3	9.5	9.5	11.4	10.1	10.2	10.2	10.1	8.8	9.0	9.0
MnO	.17	.26	.29	.30	.18				.18	.29	.32	.33	.19	.22	.26	.28
MgO	8.7	8.3	8.4	8.5	7.3	7.8	8.0	8.2	7.8	8.4	8.4	8.6	6.5	7.4	7.5	7.8
CaO	11.3	11.9	11.7	11.6	11.9	11.3	11.2	11.1	11.2	11.5	11.3	11.2	11.1	11.2	11.0	10.8
Na_2O	2.75	2.24	2.15	2.10	2.77	2.34	2.22	2.15	2.74	2.03	1.95	1.89	2.76	2.55	2.44	2.35
K_2O	.27	.42	.40	.36	.17	.49	.47	.41	.22	.39	.38	.34	.48	.55	.54	.48
P_2O_5	.16	.09	.06	.06	.17	.11	.07	.07	.14	.09	.06	.06	.22	.12	.09	.08
% melt		36.5	29.2	22.3		44.5	34.6	26.0		34.5	28.0	21.1		50.3	40.6	30.6
residual																
GAR		21.0	26.9	31.1		20.8	28.0	33.0		29.4	34.9	38.9		14.7	22.7	28.6
CPX		42.5	43.9	46.6		34.7	37.4	41.0		36.1	37.1	40.0		35.0	36.7	40.9

* Columns 1 to 4 are anhydrous equivalents of columns 1 to 4 in Table 10.1.
Columns ba, a, and d are the least squares best approximation of the adjacent ocean floor basalt compositions when melts with the composition of basaltic-andesite 876, andesite 66, and dacite 874, respectively, are combined with the 27kb pyroxene and garnet compositions listed in Table 10.2, in the proportions indicated.

** Total Fe as FeO.

Estimates of the source composition generally lie within the range indicated for oceanic layers 2 and 3 in Table 10.1 except for Ti, Mn, P, and Na. K_2O contents of 0.4 to 0.6% are not unreasonable for altered ocean floor basalts. The Na discrepancy, implying that melting under these conditions will yield rocks with several per cent more Na_2O than observed in Namosi or any other island arc andesites, is more severe and would require melting at greater depths where residual pyroxene will be richer in jadeite. Otherwise it seems possible to produce rocks with the major element compositions of the Namosi Andesite suite by high degrees of partial melting of eclogite (Green and Ringwood, 1968, envisaged 40 to 50% melting) below 80 km depths. Strictly this applies only to the andesite composition and then only when it contains 5% water. This may approximate conditions at the amphibolite-eclogite boundary but cannot apply to dry eclogite melting as it requires ∿1.5% water in the source rock. Although it was not possible to assess the higher pressure condition of dry melting where andesite is the low-melting fraction, this mathematical exercise suggests that it is, in principle, possible.

10.3 Eclogite melting and Namosi Andesites: trace elements and Sr isotopes

Shaw (1970, equations 14 and 15) provides an algebraic basis for calculating trace element behavior during non-modal partial melting for conditions when the liquid is in either surface or bulk equilibrium with the residue. If one of these equilibrium conditions prevailed and one knew the original subsolidus mineralogy and trace element contents of the source, the degree of fusion and percentage of each phase melted, and the appropriate distribution coefficients, then the trace element composition of the resulting liquid could be calculated. Estimates of trace element concentrations in ocean floor basalts are given in Table 10.1, and estimates for clinopyroxene/liquid and clinopyroxene/garnet distribution coefficients (based on data for phenocryst/lava pairs and minerals separated from eclogites, respectively) in Table 10.4. Degrees of fusion calculated as discussed in the preceding section are given in Table 10.3.

As a limiting case I calculated the concentration of several trace elements in hypothetical liquids formed by partial melting of eclogite when only pyroxene and garnet occur as residual phases. Results summarized in Figure 10.1 illustrate the range of predicted concentrations in magmas produced by either batch or Rayleigh melting

Table 10.4. Distribution coefficients used.**

	1. L/Cpx	range	2. Cpx/Gar	range	3. Hb/L	4. Mag/L	5. Pl/L
Rb	17.	-	2.		0.06*	-	0.04
Sr	6.7	-	20.		0.25*	-	1.5
Ba	20.	-	2.		0.25*	-	0.16
La	5.	2.5-10.	40.	20-60	(0.10)	-	(0.2)
Ce	4.	1.7-10.	20.	3-33	0.15	-	0.2
Pr	3.	1.6-10.	10.	2.3-18	0.22	-	0.17
Nd	2.5	1.4-10.	7.	4-10	0.3	-	0.15
Sm	1.8	1.2-10.	2.	1.3-2.5	0.4	-	0.13
Eu	1.65	1.1-10	1.3	0.8-1.4	0.5	-	0.25
Tb	1.6	1-2.	0.32	0.2-0.5	-	-	-
Ho	1.6	1-2.	0.14	0.1-0.2	-	-	-
Er	1.65	0.8-9.	0.10	0.08-0.11	0.6	-	0.4
Yb	1.75	0.8-10.	0.05	0.03-0.07	0.5	-	0.3
Y	1.7		0.10		(0.6)	-	(0.5)
Ni	0.33		5.		2.8*	8.	-
Co	0.5		0.5		3.*	10.	-
Cr	0.03		1.5		1.3*	50.	-
Sc	0.33		0.4		2.2*	2.	-
V	1.0		5.0		2.5*	40.	-

Notes:

*Calculated from phenocryst/liquid data for Namosi andesite 66 to be found in Tables 4.3, 4.4, and 10.2. They agree with those found in the literature.

**Data sources as follows:

Column 1: Schnetzler and Philpotts (1970)
Philpotts and Schnetzler (1970)
Gast (1968)
Griffin and Murthy (1969)
Hakli and Wright (1967)
Masuda and Kushiro (1970)
Ewart (pers. comm., 1971) for Tongan basaltic-andesites
Jakeš (1970) for Mau Quarry andesite

Column 2: above sources plus
Masuda (1967)
Jakeš (1970) for Kakanui eclogite
Taylor, Kaye, et al. (1969)
Coleman et al. (1965)
Bakun-Czubarow (1971)

Column 3: above sources.
Note very different values for dacites in Nagasawa and Schnetzler (1971).

Column 4: Duncan and Taylor (1968)
Ewart (pers. comm., 1971) for Tongan samples

Column 5: above sources plus
Korringa and Noble (1970)
Rb, Sr, Ba, and Eu values were chosen to apply to anorthite compositions

Most crystal/liquid data are based on phenocryst/lava pairs and are thus subject to uncertainties about equilibrium and effects of pressure. See above sources plus Albarede and Bottinga (1972) for discussion.

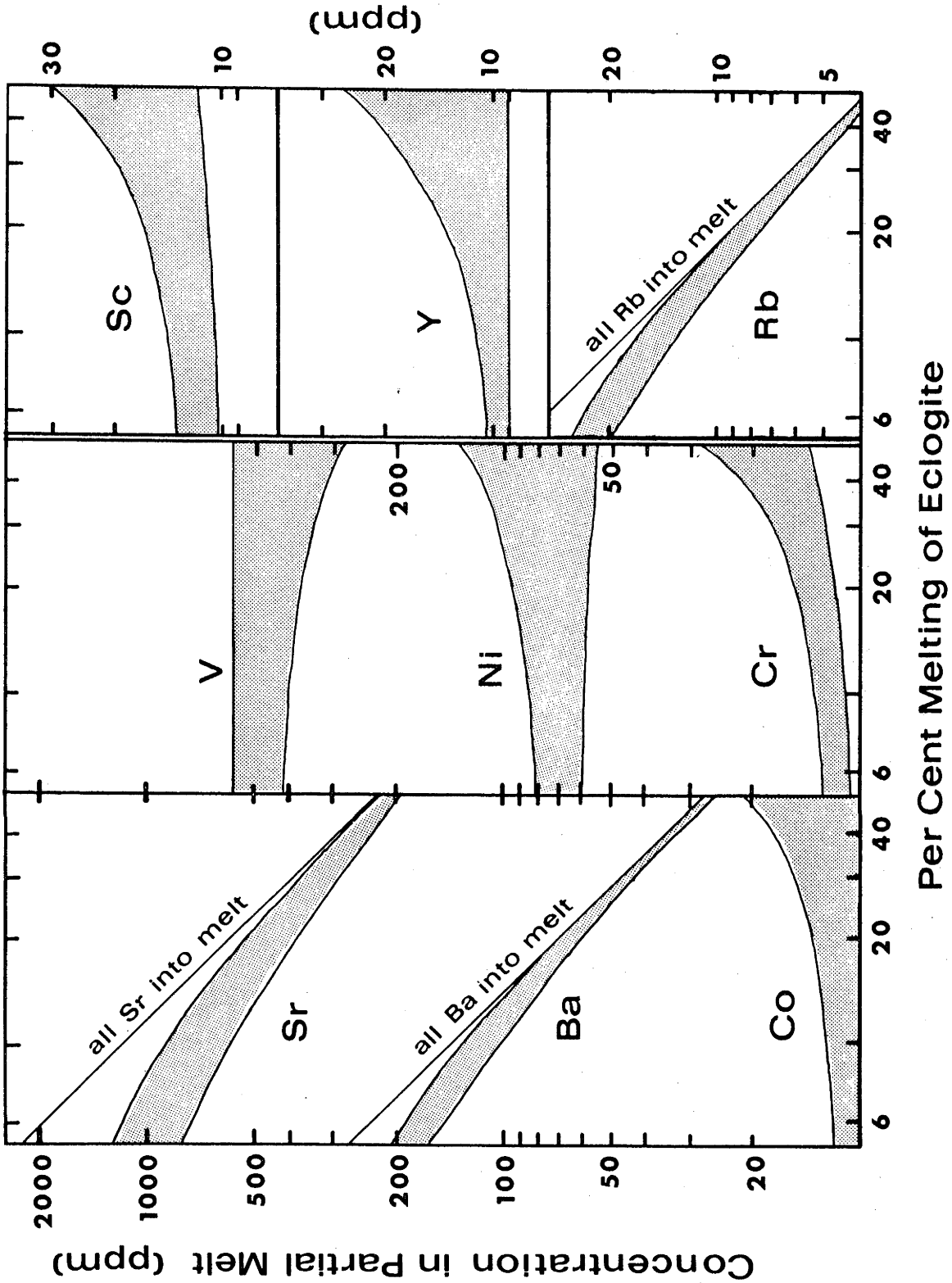

Figure 10.1 Trace element concentrations in melts formed by partial melting of eclogite. Under conditions discussed in text. Shaded regions indicate the range of expected concentrations in magmas resulting from 5 to 50% Rayleigh or batch melting in which the ratio of garnet to pyroxene being melted is between 0.6 and 4.0. The subsolidus assemblage is assumed to be .475 garnet + .475 pyroxene + .05 soluble accessory minerals and to have had the initial trace element contents given in Table 10.1, column 7. Distribution coefficients used are listed in Table 10.4.

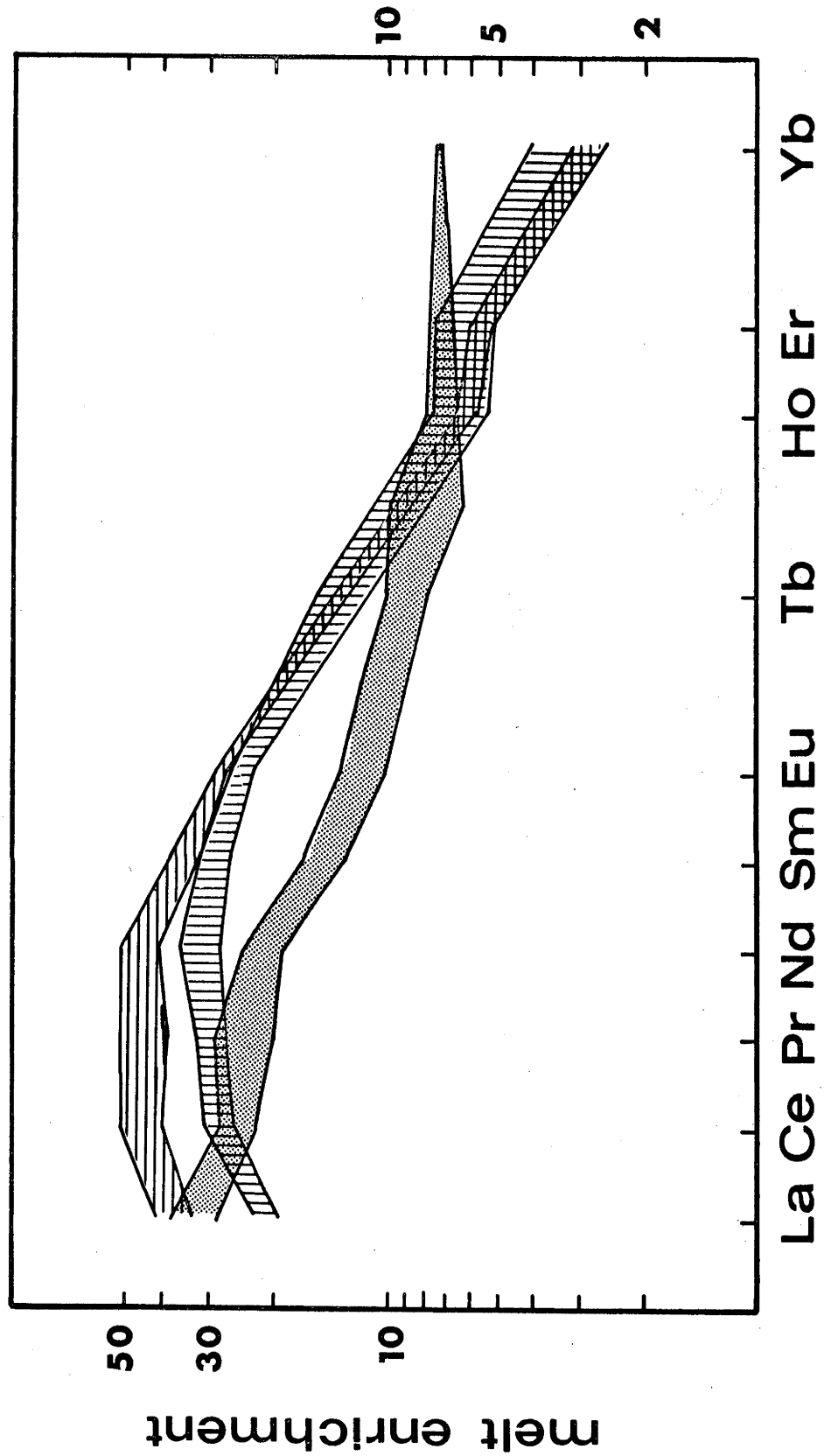

Figure 10.2 REE concentrations in melts formed by partial melting of eclogite. Assumptions are the same as for Figure 10.1. The initial concentration was taken from Schilling (1971, Table 1). The vertically striped region indicates the range of concentrations predicted for 20% melting; the horizontally striped region is the same for 40% melting. Results of 30% melting are intermediate between these regions. The dotted area indicates the observed range of Namosi Andesite REE concentrations taken from Figure 4.5.

if the initial subsolidus assemblage contained 47.5% clinopyroxene + 47.5% garnet + 5% accessory phases which completely dissolved in the liquid (what phases these might be is immaterial if they completely dissolve: see algebra in Shaw [1970] or Gast [1968]), and if the ratio of garnet to pyroxene being melted varied from 0.6 to 4.0. Similar curves were constructed for other initial assemblages. They have identical shapes but occur at slightly different concentration levels depending on the difference between pyroxene and garnet distribution coefficients and the amount of soluble accessory minerals initially present.

These calculations are very assumption-dependent but illustrate several things about Namosi Andesites relative to eclogite melting under these conditions. (1) Fresh ocean floor basalts have too little K, Rb, Sr, and especially Ba. Altered basalts can satisfy K, Rb, and Sr content and isotope ratio requirements but leave Ba as anomalous, regardless of assumptions about subsolidus mineralogy or melting behavior. Pb, Th, and U contents and K/U and Th/U ratios are consistent with an altered but not fresh ocean floor basalt ancestry. U but not Th contents increase and Th/U ratios decrease in more acidic Namosi rocks, which is not expected a priori from progressive melting of eclogite. (2) Predicted Co, Sc, and Cr contents are close to those of Namosi Andesites whereas V and especially Ni contents are too high. The Ni problem was also noticed by Taylor et al. (1969) and Hedge and Lewis (1971) in other areas. The observed rate of concentration change is generally twice the maximum predicted for each of these elements except Ni. Although the predicted rate of change between 20 and 45% melting depends on the kind of equilibrium attained, the relative amounts of minerals melted, and the absolute distribution coefficients, the differences in Co, Cr, Sc, and V contents between basaltic-andesite and dacite is inconsistent with eclogite melting behavior as constrained by data in Table 10.3. (3) As noted in the preceding subsection, Ti values predicted for Namosi Andesites are anomalously high. Their Zr contents (99 $\pm$14) overlap those of ocean floor basalts (Cann, 1970), but their Hf contents are lower (compare with data of Brooks, 1970) resulting in higher Zr/Hf ratios (70 to 95 rather than 28 to 51). The positive Ti-Zr correlation of ocean floor basalts ($r = .84$: Pearce and Cann, 1971) becomes negative in Namosi Andesites ($r = -.74$; significant at 97.5% confidence level). (4) Sn and Nb contents are low relative to those of most basalts or

andesites (see Hammaguchi and Kuroda, 1969; Parker and Fleischer, 1968; and Vlasov, 1966). (5) Using the average REE contents given by Schilling (1971) for oceanic tholeiites (= ocean floor basalts) REE patterns were calculated for various partial melting cases. Those shown in Figure 10.2 are for 20, 30 and 40% melting under the same conditions which were described at the beginning of this subsection. Higher initial garnet/pyroxene ratios result in less light and more heavy REE in the melt. The four REE patterns of Namosi Andesites are reproduced in Figure 10.2 for comparison. Predicted melts have more Ce to Tb, less Yb and Y, and greater variability in heavy REE concentrations than those of the Namosi suite. To assess how assumption-bound the disagreement was, a computer search was made using the range of ocean floor basalt REE contents given by Schilling (1971) and the range of distribution coefficients given in Table 10.4. Degrees of melting were specified as 20, 30 and 40%; the initial ratio of garnet to pyroxene was allowed to vary from 0.5 to 1.0; the ratio of garnet to pyroxene being melted from 0.6 to 4.0; and the amount of soluble accessory phases from 0 to 15%. Predicted concentrations never came within 5% of measured abundances for all rare earth elements at the same conditions.

These five conclusions are premised on so many hazardous assumptions that they cannot exclude Green's proposal, but they at least demonstrate that the simplest most straightforward version of it is inadequate. A combination of differences due to melting behavior of accessory minerals, involvement of oceanic sediments, disequilibrium conditions, and radically different distribution coefficients are needed to resolve the diverse discrepancies outlined. Assessment of the latter will require high pressure experimental study; the first two possibilities are discussed in the following sections.

10.4 Effects of ocean floor sediments

Armstrong (1971) and others have argued that most island arc rocks have too much K, Rb, Cs, Pb, Th, U, and Ba to represent partial melts of fresh ocean floor basalt alone and therefore speculate that oceanic sediments remain part of underthrust lithosphere and contribute to partial melts of it. The "average ocean floor sediment" composition given in Table 10.1 is from Wedepohl (1969) and Turekian and Wedepohl (1961), is subject to wide variability (see Wolf *et al.*. 1967; Thompson, 1968; Bender and Schultz, 1969), but is nevertheless used in the discussion which follows as a first approximation. Incorporation of such sediment could eliminate the Ba and other large cation anomalies

but, in so doing, would increase the discrepancy between predicted and observed Na, P, Ni, and light REE concentrations and Zr/Hf ratios.

If fresh ocean floor basalt supplemented by oceanic sediment is envisaged as being parental to island arc magmas, then the difference between 87Sr/86Sr ratios of such basalts (0.7025: Hedge and Peterman, 1970) and of, for example, Namosi Andesites (0.7037) must be explained by the sediment contribution. Church (1971), using these same ideas and figures to evaluate sediment contribution to Cascade andesitic volcanism, calculated that 10% of the source was sediment with an 87Sr/86Sr ratio of 0.709. If all Sr partitions into the melt (a close approximation: Figure 10.1), a 33% partial melt (estimated for a Namosi andesite from Table 10.5) of 90% fresh ocean floor basalt plus 10% oceanic sediment will have Sr with the appropriate 87/86 ratio of 0.7037, but only 400 ppm of it. This is too low for Namosi (545 $\pm$67 ppm) or Cascade andesite ($\sim$700 ppm). Higher 87/86 ratios or lower Sr contents in clays compound the problem. Approximately 33% partial fusion of a mixture of 4.5% ocean floor sediment + 3% carbonate (190 ppm Ba, 2000 ppm Sr with an 87/86 ratio of 0.709) + 92.5% fresh ocean floor basalt during which all Sr and Ba partitions into the melt could provide a magma with the Ba and Sr contents of Namosi Andesites. Such Sr would have an 87/86 ratio of 0.7030 if that of the sediment were 0.709, or 0.7035 if that of the sediment were 0.72, a less likely alternative for western Pacific oceanic clays. Preservation of sediment with a 4.5:3 clay to carbonate ratio requires special circumstances insofar as western Pacific oceanic crust will have spent $>10^7$ years below the carbonate compensation depth. Water and carbonate-rich sediments will partially melt if temperatures exceed even 600°C during underthrusting and are thus anticipated to lose their Sr and Ba before reaching the eclogite stability field.

Therefore, oceanic sediments were probably not involved in the genesis of the Namosi Andesites under the conditions being considered.

10.5 Effects of accessory minerals

Important assumptions of the calculations discussed in sections 10.2 and 10.3 were that pyroxene and garnet would be the only residual phases at the degrees of melting being considered. Given a fixed initial trace element content of the source and constant distribution coefficients, accessory phases will affect concentrations in subsolidus minerals but will not affect the composition of a partial

melt in which they completely dissolve. That is, the concentration of trace elements in the melt will be identical for cases 1, 2, and 3 in Chart 1 below although initial concentrations in phases A, B, and C will be different.

CHART 1

Case		Weight Fraction of Phases		
		A	B	C
1	SS	0.1	0.45	0.45
	R	0.0	0.45	0.45
2	SS	0.0	0.45	0.55
	R	0.0	0.45	0.45
3	SS	0.0	0.55	0.45
	R	0.0	0.45	0.45

SS = Subsolidus; R = Residual after 10% melting

If, however, any accessory minerals (e.g. quartz, kyanite, biotite, apatite, rutile, zircon, sulfides) are refractory, they will especially affect the minor and trace element composition of a derived magma.

Quartz and biotite will fuse quickly under dry or slightly hydrous conditions in the presence of a water-undersaturated liquid (Green and Ringwood, 1968; T. Green, 1968; Yoder and Kushiro, 1969). One to fourteen per cent kyanite is predicted in the subsolidus assemblage of ocean floor basalt (Table 10.1) depending on bulk composition, iron oxidation state, and the composition of garnet formed. The more altered the basalt, the more kyanite is likely in an eclogite facies assemblage. Kyanite persists above the solidus of incipiently melted eclogite found in kimberlite pipes (Switzer and Melson, 1969) where, when in contact with glass, it had reacted to form corundum plus mullite. It was not observed to co-exist with liquid in Green's runs, but neither was it found in subsolidus assemblages of the andesite composition of Green and Ringwood (1968) despite the presence of 8% kyanite in its eclogite norm. Kyanite can contain Cr and other trivalent ferromagnesian trace elements in octahedral sites and will therefore influence the trace element character of derived melts if not quickly and entirely dissolved.

Table 10.5. Revised approximations of source compositions, allowing for residual rutile, apatite, and mullite.*

	1.	ba	a	d	2.	ba	a	d	3.	ba	a	d	4.	ba	a	d
SiO_2	49.9	50.0	50.2	50.2	50.0	50.3	50.3	50.4	49.1	49.1	49.3	49.4	51.6	51.4	51.6	51.7
TiO_2	1.4	1.4	1.4	1.4	1.5	1.5	1.5	1.5	1.4	1.4	1.4	1.4	1.6	1.6	1.6	1.6
Al_2O_3	16.0	16.2	16.1	16.1	17.3	17.3	17.3	17.3	15.9	16.9	16.8	16.8	15.4	16.4	16.2	16.2
FeO**	9.3	9.3	9.4	9.3	8.8	8.5	8.5	8.4	11.4	10.1	10.2	10.1	10.1	8.6	8.9	8.8
MnO	.17	.25	.28	.29	.18	.22	.24	.25	.18	.29	.32	.33	.19	.22	.26	.28
MgO	8.7	8.3	8.4	8.6	7.3	7.8	7.9	8.1	7.8	8.3	8.4	8.6	6.5	7.4	7.5	7.8
CaO	11.3	11.8	11.6	11.5	11.9	11.6	11.5	11.4	11.2	11.4	11.2	11.2	11.1	11.1	10.8	10.6
Na_2O	2.75	2.23	2.15	2.10	2.77	2.35	2.26	2.20	2.74	2.02	1.94	1.89	2.76	2.54	2.43	2.34
K_2O	.27	.42	.40	.36	.17	.44	.41	.36	.22	.39	.38	.34	.48	.56	.55	.48
P_2O_5	.16	.16	.16	.16	.17	.17	.17	.17	.14	.14	.14	.14	.22	.21	.21	.21
% melt		36.6	29.2	22.3		38.8	29.3	21.9		34.1	27.7	20.8		50.7	41.1	30.9
residual																
GAR		20.7	26.1	30.2		14.6	18.5	22.4		29.2	34.6	38.6		14.2	22.2	28.2
CPX		41.9	43.7	46.5		43.2	47.9	51.3		36.0	37.0	39.8		34.0	35.6	39.8
RUT		0.6	0.6	0.6		0.7	0.8	0.8		0.5	0.6	0.6		0.8	0.8	0.9
AP		0.2	0.2	0.2		0.2	0.2	0.2		0.2	0.2	0.2		0.2	0.3	0.3
MUL			0.2	0.2		2.5	3.4	3.5								

* Format and calculations as in Table 10.3. The following compositions were used for minor residual phases: rutile (100% TiO_2); apatite (56.8% CaO + 43.2% P_2O_5); and mullite (28.2% SiO_2 + 71.8% Al_2O_3).

** Total Fe as FeO

Rutile (1 to 2%) and apatite (<0.5%) are also expected in subsolidus assemblages (Table 10.1). The calculations described in section 10.2 and summarized in Table 10.3 were repeated, making provision for the possibility of refractory rutile, apatite, and mullite. These results are in Table 10.5. Approximation to ocean floor basalt compositions, especially for Ti and P, is generally enhanced. Unless rutile and apatite are refractory, partial melting of eclogite will yield melts with TiO_2 and P_2O_5 concentrations considerably greater than those observed in Namosi or most other island arc andesites. In addition to containing Zr, Hf, and Nb (Vlasov, 1966) which, as noted, are relatively low in Namosi Andesites, rutile can also accept trivalent ferromagnesian trace elements despite charge imbalance problems. Apatite can contain significant amounts of Sr and large and/or highly charged cations including REE and is reported to occur as inclusions in eclogite garnets. Either of these phases could play an important role in trace element distribution if not quickly dissolved during partial melting. Refractory zircon would likewise affect Zr, Hf, P, and REE contents.

Finally, sulfides including pyrrhotite-pentlandite$_{ss}$ are found in some eclogites (e.g. Meyer and Brookins, 1971). Naldrett (1969) and MacLean (1969) report that pyrrhotite can remain stable above basalt liquidi depending on Fe^{+2} activities and show that it has a positive dP/dT slope unaffected by water content of the system or minor element content of the sulfide. Fleischer's (1955) summary of pyrrhotite geochemistry indicates the possibility of it containing up to 7.5% Ni, 0.7% Cu, 0.8% Co, and 0.3% Mn. Cambel and Jarkovsky (1969) found the concentration of these elements in pyrrhotites to increase with metamorphic grade.

10.6 Evaluation of the role of eclogite melting in the genesis of Namosi Andesites

Although the liquidus phase compositions used in the above exercise are strictly applicable only to melting conditions unrealistic for generation of a predominantly andesitic suite, they suggest it is, in principle, possible to obtain magmas with the major element composition of basaltic-andesites to dacites of the Namosi suite through 20 to 45% partial melting of eclogite having the composition of ocean floor basalt. Alkali trace element contents and isotopic ratios and Pb, Th, and U contents can be satisfied if the ocean floor basalt were

altered before subduction; sediment contamination is unlikely. Ba contents are anomalous.

Ti and P contents in such magmas will be too high unless rutile and apatite (or another phosphate) are refractory phases. Ni and V contents will also be too high and the differences between concentrations in basaltic-andesite and dacite of most ferromagnesian trace elements will be too great unless minor phases such as kyanite, rutile, or pyrrhotite$_{ss}$ dissolve gradually instead of immediately. In contrast to Namosi Andesite REE patterns, those predicted for eclogite partial melting without refractory accessory minerals retain the concave-down appearance of ocean floor basalts, have more intermediate and less heavy REE, and have greater variability between heavy REE and Y contents of basaltic-andesite, andesite, and dacite. Apatite, more enriched in light and intermediate than heavy REE (Nagasawa, 1970; Nagasawa and Schnetzler, 1971), would ease the problem if refractory; zircon would exacerbate it while solving the Zr-Hf dilemma. One or both may be necessary to explain the Th-U behavior noted.

These numerical experiments are very assumption-bound and cannot refute Green's proposal. They do not, however, give it much support. Moreover, the Namosi suite is more likely than most designated "calc-alkaline" to have evolved through eclogite melting in that it has relatively low heavy REE and Y contents. In contrast, the various average andesite types described by Taylor (1969) and the hypersthenic rock series of Japan (Taylor and White, 1966; Yajima et al., 1968) generally have ≧2 ppm Yb and >20 ppm Y. (Bougainville and New Zealand andesites [Taylor et al., 1969; Taylor and White, 1966] are similar in REE contents to Namosi ones.) Models involving eclogite will be even more difficult to reconcile with these data.

10.7 High vs. low pressure fractionation of Namosi Andesites

The preceding exercises assumed Namosi Andesites were primary, unfractionated melts. I examined two of the many other possibilities: fractional crystallization of eclogite from earlier derived andesite or basaltic-andesite; and low-pressure, relatively near-surface fractionation.

T. Green (1972) advocated the former and presented results of calculations suggesting that removal of 15% garnet and 12% clinopyroxene from andesite 66 would leave a melt with the approximate composition of dacite 874. He used the compositions of minerals

crystallizing at 36 kb, 1000 to 1100°C, in the presence of 10% water, which are given in Table 10.2 (microprobe analyses: T. Green, pers. comm., 1971). As noted in section 10.2 these minerals were not in equilibrium with liquid having the composition of andesite 66. I made similar calculations by the least squares method described above using both 36 kb sub-liquidus and 27 kb liquidus mineral compositions; results are in Table 10.6B. As with partial melting models, agreement, although not exact, is best for major elements and appears possible in principle. Garnet/pyroxene ratios for the crystallizing phases are predicted to be about 2:1 by my calculations and 1.25:1 by Green's for the 36 kb compositions, and 1.6:1 by my calculations for the 27 kb compositions. In Green's experimental runs they were ∿4:1 and ∿1:4 at 36 and 27 kb, respectively (pers. comm., 1971). Neither eclogite separation scheme removes enough Ti or K; both remove too much Mg and Mn. All such calculations are over-simplifications as they assume equilibrium rather than fractional crystallization.

Nevertheless, using the relative amounts of precipitate predicted by these calculations, I also calculated what concentrations of trace elements would occur in the resulting liquid using distribution coefficients given in Table 10.4. If bulk equilibrium between crystals and liquid is maintained, then the equation

$$C_X = C_X^o / (1 + F_X(D_X - 1) \qquad (1)$$

defines the system where C_X = concentration of element X in the derived liquid, C_X^o = concentration of element X in the initial liquid; F_X = fraction of initial mass precipitated as crystals; and D_X = bulk distribution coefficient (solid/liquid) of those crystals. If, instead, fractional crystallization involving only surface equilibrium pertains, then

$$C_X = C_X^o \cdot (1 - F_X)^{(D_X-1)}. \qquad (2)$$

Applicability of equation (2) is discussed by Neuman et al. (1954). Results of calculations using equation (2) are included in Table 10.6B; results using equation (1) are included in parentheses when significantly different.

If the distribution coefficients used are even approximately correct, eclogite crystallization will remove almost all of the heavy REE and (depending on equilibrium conditions) most of the Cr, will enrich residual liquids in V, and will remove at least half the Y. None of these effects is observed.

Table 10.6. Calculated effects of fractionation within the Namosi Andesite suite.

	A. Basaltic-Andesite from Andesite				B. Dacite from Andesite			
		2.	3.	4.	5.	6.	7.	8.
	Basaltic-Andesite 876	Andesite +Eclogite (27kb)	Andesite +Eclogite (36kb)	Andesite +Phenocrysts	Dacite 874	Andesite -Eclogite (27kb)	Andesite -Eclogite (36kb)	Andesite -Phenocrysts
SiO_2	56.16	56.62	56.55	56.27	65.74	65.29	65.49	65.74
TiO_2	0.84	0.78	0.76	0.86	0.53	0.60	0.63	0.44
Al_2O_3	18.05	17.55	17.49	17.87	16.67	16.67	16.57	16.68
FeO	7.32	7.83	7.87	7.34	3.88	4.07	3.98	3.88
MnO	0.14	0.21	0.19	0.12	0.10	0.04	0.07	
MgO	4.32	4.63	4.39	4.12	1.60	0.89	1.28	1.64
CaO	8.47	8.02	8.36	8.81	5.20	5.72	5.36	5.16
Na_2O	3.47	3.22	3.30	3.51	4.62	4.85	4.69	4.50
K_2O	1.05*	1.01	0.98	1.00	1.46	1.65	1.70	1.61
P_2O_5	0.24	0.16	0.16	0.16	0.26	0.27	0.27	0.26
	Per Cent Crystallization**							
GAR		16.0	16.9			14.4	13.6	
CPX		4.3	5.7			9.1	7.7	2.1
HB				11.6				10.8
PL				9.8				8.3
(%An)				(100)				(100)
OX				1.2				1.7
TOTAL		20.3	22.6	22.6		23.5	21.3	22.9
	Trace Element Composition***							
	Mau Quarry Andesite				(Dacite 874)			
Rb	20.	17.	18.	18.	28.	26.	25.	26.
Sr	490.	666.	684.	567.	660.	630.	614.	533.
Ba	475.	396.	408.	390.	420.	615.	599.	588.
Ni	10.	7.	7.	6.	7.	9.	9.	7.
Co	16.	13.	12.	17.	10.	9.	9.	11.
Cr	39.	(1.4) / .04	(1.3) / .02	4.	6.	(6.) / .04	(6.) / .10	(15.) / 7.
Sc	16.	7.	7.	24.	8.	4.	5.	14.
V	172.	326.	330.	(183.) / 153.	85.	196.	194.	(99.) / 74.
La	9.1	10.7	11.0	10.7	10.5	11.6	11.4	11.3
Ce	20.	22.	22.	22.	22.	25.	25.	25.
Pr	2.7	3.0	3.0	3.0	3.5	3.4	3.3	3.3
Sm	2.4	3.1	3.1	3.2	3.3	2.8	2.7	2.8
Eu	.77	.85	.86	.82	1.0	.88	.87	.84
Er	1.2	(.79) / .58	(.77) / .55	1.6	1.4	(.70) / .55	(.73) / .58	1.4
Yb	1.4	(.54) / .24	(.53) / .21	1.6	1.4	(.58) / .28	(.60) / .32	1.6
Y	18.	9.	8.	23.	18.	8.	9.	20.

*K_2O of basaltic-andesites other than 876.

**Amounts of crystals (T) which when added to (100-T)% of andesite 66 or subtracted from 100% of andesite 66, give the compositions listed above.

***Calculated as discussed in text. Those in section A refer to precipitation of the amounts of minerals listed from basaltic-andesite 876 (K, Rb, Sr, Ba data of sample 869).

A similar exercise involving addition rather than subtraction of mineral compositions was used to assess the relationship between andesite 66 and basaltic-andesite 876; results are in Table 10.6A. As found by T. Green (1972, Table 2), attempts to approximate the basaltic-andesite major element composition are less successful. (It should be borne in mind that these calculations are inherently limited by assuming equilibrium, not fractional crystallization.) The trace element composition predicted for a residual liquid remaining after crystallization of the amounts of garnet and pyroxene indicated in Table 10.6A from a melt with the initial concentrations of basaltic-andesite 876 are also included in Table 10.6A. The same discrepancies between prediction and observation that were noted above occur again.

For comparison, I repeated all these calculations using the composition of analyzed phenocrysts naturally occurring within sample 66. These are from Jakes (1970) and are listed in Table 10.2. The least squares best approximation of the dacite composition is slightly better, and of the basaltic-andesite composition considerably better than those obtained using high-pressure phases (Table 10.6). Note that no clinopyroxene was added in the closest approximation of the basaltic-andesite composition and little was subtracted in that of the dacite composition despite the common occurrence of clinopyroxene as a phenocryst, and that plagioclase involved in both simulations is 100% anorthite. (The computational method allows for separate entry and subsequent mixing of pure anorthite and albite.) Clinopyroxene was ubiquitous in Green's experimental runs, and water pressures sufficient to stabilize amphibole as a liquidus phase usually inhibited plagioclase crystallization (although they also increased its anorthite content). Jakes (1970, chapter 4) found a small field at <3 kb where both plagioclase and amphibole were liquidus phases of an andesitic melt containing 5% water, but its composition had lower Mg and Ca and higher K contents than Namosi rocks. His andesitic composition which was closer to Namosi analyses had no such field. Moreover, clinopyroxene, not amphibole, appears to have been a liquidus phase in Mau Quarry rocks on petrographic evidence (some amphibole phenocrysts have clinopyroxene cores and calcic plagioclase inclusions and all have strongly reacted rims; plagioclase phenocrysts contain inclusions of pyroxene and magnetite; those of clinopyroxene contain inclusions of plagioclase). Mau Quarry hornblendes, however, are similar in composition to those found by Green at high pressures (compare

analyses 5 and 6 in Table 10.2 with Figure 9 of T. Green, 1972). If plagioclase and amphibole were near-liquidus phases, their fractional crystallization would be consistent with the trace element compositions of Namosi Andesites as can be judged from calculations summarized in Table 10.6.

Decreasing Na/K, K/Rb, and Th/U ratios are compatible with removal of plagioclase and amphibole but not clinopyroxene. Ti behaves quite differently from Zr and Hf. This seems to require solid-liquid equilibrium with an oxide phase that accepts Ti but excludes Zr and Hf (Ti/Zr ratios decrease but Zr/Hf ratios remain relatively constant). The low pressure fractionation scheme set out in Table 10.6 implies crystallization of about 40% of basaltic-andesite melt to yield residual dacite, including crystallization of 16% of it as plagioclase. Absence of a negative Eu anomaly in dacite 874 does not preclude this much plagioclase precipitation if the plagioclase is very calcic (as Eu/Eu* ratios will approach 1: Philpotts, 1970).

No model discussed in this chapter fits all observations. In balance, however, rocks of the Namosi Andesite suite seem more related to low than high-pressure fractionation processes providing that amphibole and plagioclase can both be near-liquidus phases in some circumstances. As noted in section 4.3, hornblende-bearing members of the suite are typically younger and more silicic than hornblende-free varieties (possibly suggesting magma chamber crystal settling), and hornblende-bearing varieties have more crystalline, less glassy matrices (possibly suggesting lower temperatures of extrusion and, therefore, higher water contents). These and the other factors discussed in this chapter suggest to me that Namosi Andesite magmas acquired their primary geochemical characteristics during fractionation under high partial water pressures at fairly shallow depths where, somehow, anorthite was a near-liquidus and magnetite a minor phase. Pressure and water release (e.g. during magma ascent) may or may not render amphibole unstable (depending on temperatures, bulk composition, and volatile contents) resulting in its partial resorption and the crystallization of clinopyroxene. Some such model appears more consistent with field and petrographic evidence as well as the major and especially minor and trace element compositions of Namosi Andesites than do those models invoking crystallization or partial melting of eclogite.

10.8 High-K calc-alkaline rocks and initial comments on the cause of spatial variations in the geochemistry of island arc magmas

In this thesis the adjectival phrase "high-K calc-alkaline" is used to denote a rock series having calc-alkaline characteristics (see section 10.1) but with high concentrations of potassium and associated elements (see section 1.3). As mentioned in section 3.3, various alkali-rich rock types commonly occur late in the history of and/or furthest from the trench associated with island arcs. Sometimes these alkali-rich rocks represent a continuum between shoshonites and alkali basalts (see chapter 12). In eastern Papua and the New Guinea Highlands and apparently in Japan and Kamchatka there is also a close geochemical, spatial, and temporal relationship between shoshonitic and high-K calc-alkaline rock series which makes such naming of K-rich rocks rather arbitrary (Smith, 1970; Jakes and Smith, 1970; Peterman et al., 1970). However, the high-K calc-alkaline and shoshonitic rock series do not seem to be closely related in Fiji. Calc-alkaline rocks associated with shoshonites in Fiji (section 4.4E) are not high-K by my definition (section 1.3). High-K calc-alkaline rocks in the Kadavu Group (section 7.3) are not known to be associated with shoshonites although the geology is not well enough known to exclude this possibility. The association is also possible on Beqa or Ono-i-Lau.

It is instructive to note differences between the Kadavu Group high-K andesite and first, Fijian shoshonites and second, the Namosi Andesites. Relative to shoshonitic rocks of similar potassium contents, the Kadavu sample: (1) is intermediate rather than basic with correspondingly lower ferromagnesian trace element contents; (2) has approximately similar Rb, Sr, Ba, and Pb concentrations but a higher Na/K ratio; (3) has a much lower 87Sr/86Sr ratio; (4) has considerably more Th and U with a similar Th/U but much lower K/U ratio; (5) has considerably more light REE resulting in a higher La/Yb and lower K/La ratio; and (6) has less P but more Zr, Hf, and Sn. The high concentration of K, Rb, Sr, Ba, and Pb in both rock types suggests that these elements were acquired through interaction with crustal or mantle wall rocks, and that this process is responsible for producing shoshonite from basalt and high-K from normal calc-alkaline rocks. This is consistent with the observation (section 9.2) that shoshonites and high-K andesites are often confined to areas which have thicker, older crust than do other island arcs. The dramatic differences in P,

Th, U, and light REE contents and 87Sr/86Sr ratios between these rock types in Fiji, however, excludes the possibility that the same process yielded the K, Rb, Sr, Ba, and Pb enrichments in both. Shoshonites are discussed further in section 12.2.

Except for large cation and Zr, Hf, and Sn contents, Kadavu and Namosi andesites have similar compositions. Thus, Kadavu Group andesites may represent magmas similar in composition to Namosi Andesites except for having become (more) contaminated by interaction with crust or mantle wall rocks during ascent and crystallization. Because Rb/Sr ratios are similar in both andesites (0.038) but 87Sr/86Sr ratios are lower in Kadavu, the 2.5 to 3-fold enrichment of these elements (plus Ba, Pb, Th, U, and La) in Kadavu relative to Namosi samples, must, if contamination is the cause, indicate a large cation source with a Rb/Sr ratio of 0.038 yet an 87Sr/86Sr ratio of 0.7028. This is the Rb/Sr ratio thought to characterize never-before-depleted upper mantle, suggesting a wall-rock reaction process, but such Sr would have had an 87/86 ratio $\sim$0.706 in the late Tertiary (Armstrong, 1968; Gill and Compston, Appendix 3). The Rb and Sr cannot, therefore, have been acquired from a source (such as never-before-depleted upper mantle) which has remained a closed system for 4.5 b.y., and very special conditions must be postulated if mantle wall-rock reaction is the source of the large cation enrichments.

Instead, the introduced Sr might previously have moved from a low Rb/Sr environment to one where Rb/Sr was $\sim$0.04. Qualitatively this suggests incorporation into the crust. Crust with Rb/Sr ratios $\sim$0.04 and 87Sr/86Sr ratios $\sim$0.7028 could conceivably exist beneath Kadavu but no known examples of such material have high Rb and Sr concentrations. Two to three-fold enrichments of large cations in ascending magmas would therefore require extensive crustal assimilation. Because it cannot be quantified this possibility is difficult to rule out unless the argument developed in section 9.3 is adopted and Tholo quartz-tonalites are viewed as representing the low-melting fraction of lower Lau-Colville Ridge crust. If so, crustal assimilation will not produce Kadavu sample 390 from any Namosi Andesite composition. Highly selective and efficient leaching of large cations plus Zr, Hf, and Sn (but not Cu) is another alternative which is even more difficult to evaluate. As a result I cannot, on the basis of my one sample, exclude the possibility that its enrichments are due to interaction with the crust or mantle, but I think it unlikely.

Alternatively, Kadavu Group and Namosi andesites might represent different residual liquids of the same parent. About 70% of the mass of a magma having the initial composition of a Namosi basaltic-andesite would have to precipitate as crystals containing no large cations before a residual liquid could attain the trace element composition of Kadavu Group andesite 390. Such a residual liquid would probably be rhyolitic instead of andesitic.

Finally, these differences may reflect degree of partial melting or the subsolidus mineralogy of the source which gave rise to the respective magmas or their parents. Jakes and White (1970) and Jakes and Smith (1970) suggest minor melting of mica-bearing eclogite as a source of high-K calc-alkaline rocks, but this, by itself, would lead to high Ba/Sr ratios ($\geqq 10$) in contrast to what is observed ($\leqq 1$). Moreover, the difficulties with eclogite melting models noted in section 10.2 to 10.6 apply equally to the composition of sample 390. Least squares calculations such as those described in section 10.2 suggest that the degree of melting necessary to match the major element composition of sample 390 is similar to the amount necessary to match Namosi sample 66. The trace element problems are as or even more intractable. The differences in $^{87}Sr/^{86}Sr$ ratios, if primary, require derivation of magmas with similar Rb/Sr ratios from sources which (once) had different Rb/Sr ratios. Alternatively, if Namosi andesites and dacites resulted from relatively low pressure fractionation of a basaltic-andesite or more basic parent, then perhaps the Kadavu andesite did also, with differences in large cation, Zr, Hf, and Sn concentrations and Sr isotopic character being inherited.

These ambiguities prevent the foregoing discussion from shedding much light on the source of element variation in island arcs. Element enrichments in at least the Fijian shoshonites, alkali basalts, and high-K calc-alkaline andesites must reflect different processes. In Fiji, high-K andesites appear temporally and spatially related to normal island arc volcanic activity whereas shoshonites and alkali basalts appear related to initiation and development of inter-arc rifting. Differences between regular and high-K calc-alkaline magmas might reflect (greater) interaction of the latter with crustal or mantle wall rocks but this requires very special conditions. There is no reason to expect differences in age or composition of crust between Kadavu and Viti Levu; if anything, crust appears thicker beneath southeastern

Viti Levu than beneath Kadavu (compare Robertson, 1967, with Robertson, 1965). Alternatively, differences could have been inherited from parental magmas at least as basic as basaltic-andesite.

Thus, there is little convincing evidence that element variations have a direct relationship with melting of only underthrust lithosphere. This will be considered again in the following chapter.

Chapter 11. The Island Arc Tholeiitic Series Re-examined and the Origin of Spatial Variations in the Geochemistry of Island Arc Magmas

11.1 Denotations

Comparison of tables and figures delineating the composition of first and second period Viti Levu volcanics, the Mathuandrove Super-Group of Vanua Levu, and the east and west Lau Volcanics, illustrate the diversity of rock types in island arcs to which the adjective "calc-alkaline" is often uniformly applied. Many of the differences are summarized in the "average" compositions compiled for these series in Table 13.1, and in the more general tables and figures of chapters 1 and 3. In this sense, the Namosi and Kadavu andesites discussed in chapter 10 are simply the most calc-alkaline in Fiji. First period Viti Levu volcanics are the least calc-alkaline and most tholeiitic; the Mathuandrove Super-Group and two Lau groups are intermediate between them. Jakes and I (1970), when defining the island arc tholeiitic series, stressed that it and the calc-alkaline series were end members of a continuum but had consistent compositional, spatial, and temporal relationships with each other. As discussed in section 3.4, the temporal relationship may be fortuitous; the others do not seem to be. The continuity of the relationship suggests a genetic link; the consistency suggests that the link is fundamental, recurring in most island arcs and related to underthrusting.

The idealized picture presented in Figure 3.1 is, as can be judged from data in Table 3.1, an over-simplification of modern island arcs. No one of the rock types named is ubiquitous; all three are rarely developed in the same region on a time scale of 10^6 years. But, as documented in chapter 3, where variation does occur it follows the pattern I outlined and the following discussion assumes the general validity of the model presented in Figures 3.1 to 3.3.

11.2 Comparison between the calc-alkaline (CAS) and island arc tholeiitic (IATS) series

Magmas having more tholeiitic characteristics than Namosi Andesites occur closer to the trench and above shallower earthquake foci than those which do not. Nevertheless, as noted in chapter 9 and as can be inferred by conjoining data in Table 3.1 with data in Isacks and Molnar (1971), even these magmas rarely erupt less than 100 km above Benioff zone seismicity. Eclogite rather than amphibolite will probably, therefore, be the stable assemblage in uppermost lithosphere

immediately beneath volcanoes yielding either the CAS or IATS. IATS magmas, however, are at least closer to areas in which vertically rising melts could ascend from depths at which amphibole breaks down in water-deficient basaltic compositions.

Mineralogically the more tholeiitic series usually lack hornblende and biotite, may contain groundmass pigeonite, and often contain fewer total phenocrysts. Both are dominated by zoned clinopyroxene and labradorite to bytownite phenocrysts and can contain orthopyroxene and iron oxide(s).

IATS rocks usually have a wide range of SiO_2 contents (50 to 75%) which may (as in Fiji) or may not (see examples in Table 3.1) exceed the range of calc-alkaline varieties, and they usually have a lower silica mode (i.e. basaltic-andesite rather than andesite) than do the latter. Relative to CAS rocks, those of the IATS are more likely to exhibit Fe- and Ti-enrichments (up to 16%ΣFeO at Raoul Island, Kermadec: Brothers and Searle, 1970) and to have lower Fe_2O_3/FeO ratios, resulting in more normative pyroxene at a given silica content. This propensity for Fe-enrichment, although common, is not uniform.

Most petrologists believe rocks should be classified on the basis of modal mineralogy or at least major element composition. However, the most dramatic differences between the CAS and IATS are in minor and trace element contents (see Jakes and Gill, 1970). IATS rocks have lower K, P, Rb, Ba, Sr, Cs, Pb, Th, and U contents, higher Na/K, K/Rb, K/U, and Ba/Rb ratios, and lower Rb/Sr, U/Pb, and Th/U ratios (Table 1.1). These features are especially plain in what I shall call "low-K rhyodacites" of the island arc tholeiitic series. Analyses of such material from most circum-Pacific island arcs are compiled in Table 11.1. They differ substantially in the above-mentioned element contents and ratios from most granites, ignimbrites, pantellerites, and other rhyolites and dacites. These minor and trace element features are, therefore, not ephemeral peculiarities subordinate to more important aspects of modal mineralogy or major element composition. They are distinctive and persistent, and characterize entire suites ranging from basalt to rhyolite.

IATS rocks also have different REE patterns (see Figure 4.5 and Jakes and Gill, Figure 6). They are flatter relative to chondritic abundances (i.e. have La/Yb ratios <3), may be relatively enriched in intermediate REE, and have consistently higher concentrations of heavy

Table 11.1. Analyses of low-K rhyodacites of the island arc tholeiitic series.

	A. FIJI								B. TONGA		C. KERMADEC	D. SAIPAN		E. JAPAN		
	856	387	S0	854	H26	B535	M96	AN8[1]	F39[2]	Metis Shoal Glass[3]	13[4]	139[5]	317[6]	5[13]	23[7]	11[8]
SiO_2	70.7	72.1	73.3	74.3	75.1	77.8	79.4	77.9	65.3	73.4	67.0	79.2	79.0	73.0	67.4	75.8
TiO_2	0.4	0.4	0.5	0.4	0.5	0.3	0.2	0.3	0.6	0.4	1.2	0.2	0.2	0.3	0.7	0.5
Al_2O_3	15.4	14.6	13.1	13.9	13.5	11.4	10.4	12.5	14.1	12.7	13.6	11.1	9.9	12.5	15.3	11.4
Fe_2O_3*	3.9	2.9	1.6	2.3	2.5	1.7	2.5	1.0	6.8	1.1	1.9	0.5	1.2	2.0	1.1	1.4
FeO			2.1		0.1	-	-	0.4	1.1	2.8	4.3	0.9	0.2	2.1	3.9	1.3
MnO	0.1	0.1	0.1	0.1	-	-	-	-	0.2	0.1	0.2	-	-	0.1	-	-
MgO	0.8	1.2	0,8	0.5	0.3	0.8	0.3	0.1	1.5	0.8	2.0	0.4	-	1.0	1.2	0.5
CaO	2.4	1.2	2.6	0.6	2.4	2.5	1.8	2.5	5.9	3.8	4.5	2.1	0.6	3.3	4.5	2.4
Na_2O	5.3	7.1	4.8	6.5	4.7	4.0	4.1	4.4	3.1	3.1	3.8	3.4	3.9	3.7	4.7	4.1
K_2O	0.9	0.3	0.9	1.4	0.6	1.4	1.2	0.7	1.1	1.5	0.8	1.6	1.2	1.1	0.9	1.4
P_2O_5	0.1	0.1	0.1	0.1	0.1	0.1	0.2	0.1	0.2	0.1	0.2	0.1	0.1	0.3	0.2	0.2
Rb	6.	3.	7.	15.	8.	12.	6.	-	17.	21.	-	15.	-	-	-	-
Sr	200.	90.	120.	120.	110.	90.	100.	-	300.	130.	-	90.	60.	-	-	-
Ba	210.	55.	220.	300.	90.	140.	120.	-	190.	330.	-	180.	-	-	-	-
Pb	2.	2.	1.	2.	1.	4.	3.	-	4.	-	-	4.	-	-	-	7.
Th	0.8	-	-	2,5	0.4	-	0.4	-		-	-	1.6	-	-	-	0.7
U	0.5	-	0.2	1.0	0.4	-	0.5	-	-	-	-	0.8	-	-	-	0.4
La	-	-	3.	-	61.	-	7.	-	6.	3.	-	6.	-	-	-	-
Yb	-	-	6.	-	28.	-	3.	-	2.	2.	-	2.	-	-	-	-
Y	47.	44.	44.	70.	560.	34.	52.	-	24.	25.	-	23.	-	-	-	-
Zr	160.	190.	70.	220.	180.	140.	120.	-	50.	70.	-	125.	-	-	-	-
87Sr/	-	.7045	.7047	-	-	.7043	.7042	-	.7043	.7036	-	-	-	-	-	-
86Sr	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
K/Rb	1100.	700.	1100.	800.	650.	1000.	1500.	-	550.	580..	-	870.	-	-	-	-
Rb/Sr	.031	.036	.059	.127	.075	.130	.064	-	.057	.161	-	.167	-	-	-	-
La/Yb	-	-	0.4	-	2.2	-	2.4	-	3.7	1.2	-	2.9	-	-	-	-
Zr/Hf	-	-	16.	-	79.	-	58.	-	-	-	-	48.	-	-	-	-

Notes: *If only Fe_2O_3 values are given it indicates total Fe as Fe_2O_3

References:

unnumbered = this thesis

1. Rodda (1969)
2. Ewart,(pers. comm.)
3. Melson et al. (1970) and Ewart (pers. comm.)
4. Brothers and Searle (1970)
5. Taylor et al. (1969)
6. Schmidt (1957) and Pushkar (1968)
7. Kuno (1950)
8. Kuno (1962) and Tatsumoto and Knight (1969)
9. Donnelly et al. (1971)
10. Baker (pers. comm.)
11. Bryan (1968)
12. Gass et al. (1963)
13. Ono (1962); contains many examples

Table 11.1. Continued....

	F. ANTILLES	G. SCOTIA	H. NEW BRITAIN (North Coast)		I. PUMICE	
	Average Water Island Fm. Keratophyre[9]	34[10]	30[2]	36F[2]	Coral Sea[11]	Scotia[12]
SiO_2	74.4	73.0	69.5	75.2	73.5	73.0
TiO_2	0.3	0.3	0.4	0.3	0.6	0.3
Al_2O_3	12.7	13.6	14.7	12.8	11.9	13.6
Fe_2O_3	2.1	0.7	1.7	0.8	0.8	0.7
FeO	-	2.1	2.7	1.2	4.8	2.1
MnO	-	0.1	0.1	0.1	-	0.1
Mgo	1.1	0.5	1.1	0.3	0.6	0.5
Cao	1.1	3.2	4.4	2.3	3.8	3.2
Na_2O	4.5	4.7	4.1	4.5	2.8	4.7
K_2O	0.8	0.7	1.0	1.3	0.8	0.7
P_2O_5	-	0.1	0.2	0.1	0.1	0.1
Rb	-	21.	-	-	-	-
Sr	-	160.	250.	250.	-	-
Ba	-	125.	190.	240	-	-
Pb	-	-	-	-	-	-
Th	0.7	-	0.4	0.8	-	-
U	0.5	0.6	0.4	0.3	-	-
La	-	-	-	-	-	-
Y	-	-	16.	20.	-	-
Zr	-	80.	80.	100.	-	-

REE and Y especially in siliceous examples. Negative Eu anomalies may appear in the low-K rhyolites.

Jakes and I (1970 and Gill, 1970) implied that REE patterns of the IATS and CAS were mutually exclusive. Yajima et al. (1968) found those of Kuno's pigeonitic and hypersthenic rock series to be different, that the former were comparable to our island arc tholeiitic field, and that the latter were different from but transitional to our calc-alkaline field (compare Jakes and Gill, 1970, Figure 6 with my summary of Yajima et al.'s data in Figure 11.1). REE patterns of first period Viti Levu and modern Tongan rocks fall within the "island arc tholeiitic field". REE patterns of Vanua Levu and Lau rocks are similar to those of the Japanese hypersthenic rock series (compare Figures 11.1 and 11.2) and high-alumina basalts of the Hakone area (Masuda, 1966, 1967; Philpotts et al., 1971). I shall call this a "transitional field". Namosi Andesites have comparable light but lower heavy REE and Y concentrations which remain constant rather than increasing in siliceous members. I shall confine the "calc-alkaline field" to REE patterns of this type. These three fields--tholeiitic, transitional, and calc-alkaline--illustrate the continuity and range between REE in island arc volcanic rocks.

Within tholeiitic and transitional fields there appears to be no strict correlation between REE enrichment and silica contents or Fe/Mg ratios. Both light and heavy REE contents generally increase in more siliceous rocks but not consistently. There are suggestions of concave-down patterns in the tholeiitic field and concave-up patterns in the transitional field but data are too imprecise to be confident.

Variations in ferromagnesian trace element contents are difficult to assess because they are very sensitive to the major element compositions of rocks being compared. There is, however, at least a suggestion in my data that Ni and Cr contents are lower and Sc higher in IATS than CAS rocks of comparable MgO contents (i.e. that Mg/Ni and Mg/Cr ratios are higher and Mg/Sc ratios lower in the IATS).

For a specific Rb/Sr ratio, 87Sr/86Sr ratios in IATS rocks seem to be higher than those in the CAS. Moreover, data reported by Gill and Compston (Appendix 3) are sufficiently precise to indicate inhomogeneity of Sr isotopes within some IATS suites, where they are positively correlated with Rb/Sr ratios. No such variation was found for Pb isotopes in Tongan samples (Oversby and Ewart, pers. comm., 1972)

or for Sr isotopes of less tholeiitic suites in Fiji, New Britain (Peterman et al., 1970a), the Cascades (Church and Tilton, in prep.; Peterman et al., 1970b), or the Antilles (Hedge and Lewis, 1971).

After some initial enrichment in IATS suites, Ti contents decrease in both rock series; Zr contents increase in both. Ti/Zr ratios may be higher in intermediate members of the IATS than CAS; Zr/Hf ratios appear similar (60 to 80) in both.

11.3 Comparison between the island arc tholeiitic series and ocean floor volcanic rocks

Jakes and I (1970, section 6) outlined similarities and differences between island arc and ocean floor basalts. Subsequently Hart et al. (1970) emphasized their continuity in terms of K, Rb, Cs, Sr, and Ba concentrations; Pearce and Cann (1971) noted differences in Ti, Zr, and Y behavior between them.

Most ocean floor volcanic rocks are olivine tholeiites although there is a broad continuum between ne- and hyp-normative compositions (see Aumento et al., 1971, Figure 9; Cann, 1971, Figure 2; and Kay et al., 1970, Figure 2). D. Green's (1971, 1972) proposal that these represent approximately 20 to 25% partial melts of peridotite at depths of 20 to 60 km has gained broad support. The source peridotite, however, must have been partially depleted in large cations relative to chondritic abundances before the major fusion occurred (Gast, 1968). This prior depletion probably happened during very minor partial melting (yielding highly undersaturated basalts), "wall-rock reaction", or vertical zonation of the low-velocity zone. At least the first two are convergent processes (see section 9.3). I present some criticisms of the third option, proposed by Green (1971), in Appendix 3 (Gill and Compston, section 2).

Shido, Miyashiro, and Ewing (1971) summarize evidence that ocean floor olivine tholeiites fractionate at low pressures by crystallizing plagioclase and olivine and, subsequently, augite. Kay et al. (1970) and Melson et al. (1968) also stress that plagioclase and olivine are the initial phases to crystallize; Schilling (1971) and Cann (1971) emphasize the importance of eventual clinopyroxene fractionation; and Hart (1971a) found that crystallization of these three phases plus titanomagnetite is required to produce the Amphrite 3 andesite composition discussed below. A summary of the major and trace element composition of ocean floor basalts is given in Table 10.1.

Island arc basalts are, by comparison, less common and less well studied. While tholeiitic, few basalts of pre-Pliocene Viti Levu and Vanua Levu, Fiji, or modern Japan, are ol-normative and most are qz-normative even when Fe_2O_3/FeO ratios are 0.25. This may be a primary difference or may simply reflect scarcity of little-fractionated basalt in island arcs. This uncertainty also applies to the apparently lower Ni and Cr contents of arc basalts. As emphasized by Hart et al. (1970), large cation concentrations in both basalt groups overlap but are generally lower in ocean floor varieties. The isotopic composition of Sr and possibly Pb is always more radiogenic in arc basalts. Major element compositions of arc and ocean floor basalts are compared in Table 11.2.

Differences are much more noticeable in intermediate rocks of the two areas. These rocks predominate in island arcs but are relatively uncommon in deep oceans (see review by Bonatti and Arrhenius, 1970). Analyses of intermediate rocks from the ocean floor are available from Hart (1971a), Kay et al. (1970), and Aumento (1969), and are summarized in Table 11.3. They differ most notably from analyses of arc tholeiitic andesites in their higher Ti and Fe and lower Ca and Al contents and are closer to icelandites in composition (see average icelandite analysis in Table 11.3). (Analyses of Aumento's metamorphosed "dolerites" are quite different from those of the volcanics, have very unusual Ni, Sc, and V concentrations, and are not considered further.) REE concentrations of these volcanic rocks are given by Frey et al. (1968), Philpotts et al. (1969), Haskin et al. (1970), and Kay et al. (1970), and are shown in Figure 11.1 where they are compared to schematic REE patterns for the pigeonitic and hypersthenic rock series of Hakone, Japan (Yajima et al., 1968). Schematic REE patterns for various Fijian suites are summarized in Figure 11.2. The ocean floor andesites have flat REE patterns with negative Eu anomalies as do many intermediate and acidic rocks of the island arc tholeiitic field (data from Yajima et al., 1968, for Japan; Taylor et al., 1969, for Saipan; this thesis for Fiji; Ewart and Bryan, 1972, and Hubbard in Melson et al., 1970, for Tonga). However, island arc rocks are rarely as enriched in REE and Y and then only in rhyolites, not andesites.

Ti, Zr, and Y are significantly and positively correlated in ocean floor basalts but not in the pigeonitic rock series of Japan (Pearce and Cann, 1971). They are also positively correlated in basaltic-andesites and andesites of the western Lau Volcanics

Table 11.2. Comparison of island arc, mid-ocean, and inter-arc basin tholeiite compositions.**

	1.	2.	3.	4.	5.	(1σ)	6.	(1σ)	7.
SiO_2	48.7	50.5	49.8	51.0	49.0	(0.9)	51.7	(1.4)	48-50
TiO_2	0.6	0.9	0.7	0.9	1.3	(0.5)	1.3	(0.4)	1.0-1.7
Al_2O_3	16.5	18.5	14.3	15.3	16.2	(0.7)	17.6	(2.7)	14-17
Fe_2O_3	3.4	3.4			1.9	(0.2)	1.9	(0.3)	
FeO	8.4	7.0	8.8*	9.5	7.5	(0.9)	7.6	(1.3)	6-12*
MnO	0.1	0.1	0.2	0.2	.15	(.04)	.16	(.02)	.13-.20
MgO	8.2	5.1	10.9	7.7	8.7	(2.0)	5.3	(1.3)	6-10
CaO	12.2	10.7	11.4	11.2	12.8	(1.1)	10.9	(1.0)	10.5-12.5
Na_2O	1.2	2.0	2.0	2.3	2.1	(0.5)	2.7	(0.4)	2.5-3
K_2O	0.23	0.31	0.70	0.66	0.22	(.18)	0.62	(.34)	.10-.30
P_2O_5	0.1	0.1	0.2	0.2	0.1	(.06)	0.1	(.12)	.1-.2
qz	0.4	4.3		0.3			2.0		
or	1.4	1.3	4.2	3.9	1.3		3.7		1
ab	10.2	12.5	17.1	19.6	17.8		22.9		20-25
an	39.0	37.7	28.2	29.7	34.1		34.1		25-30
di	17.3	15.0	22.1	20.3	23.2		15.9		18-20
hy	26.8	24.8	14.3	21.0	11.0		15.9		10-20
ol			9.7		7.1				2-10
mt	3.4	3.3	2.6	2.8	2.7		2.8		2
ilm	1.2	1.1	1.3	1.7	2.5		2.5		2-3
ap	0.2	0.1	0.5	0.5	0.2		0.2		0.4

*Total Fe as FeO

**Norms calculated assuming $Fe_2O_3/FeO = 0.25$.

1,2. Presumed parental magma of pigeonitic rock series, Japan; 1 from Kuno (1968b), 2 from Yagi et al. (1968).

3. Average of two Mathuandrove Super-Group olivine tholeiites.

4. Average of three Mathuandrove Super-Group quartz tholeiites.

5. Average of thirteen Lau Basin olivine tholeiites.

6. Average of six Lau Basin quartz tholeiites.

7. Range of mid-ocean olivine tholeiites (approximately one std. dev. from data in Table 10.1 and Shido et al. [1971]).

Table 11.3. Composition of intermediate rocks from the ocean floor and inter-arc basins.

	1.	2.	3.	4.	5.	6.
SiO_2	58.1	59.0	62.0	72.5	61.7	63.2
TiO_2	1.7	1.8	0.9	0.3	1.0	0.5
Al_2O_3	13.4	12.6	16.0	14.2	14.4	14.6
$Fez0_3$			3.2	1.8	4.0	3.1
FeO	10.6*	12.0*	3.6	1.2	4.3	2.5
MnO			0.1	0.1	0.2	0.1
MgO	2.9	1.7	2.4	1.4	1.0	3.4
CaO	7.1	5.6	3.2	1.5	4.3	5.2
Na_2O	4.3	4.2	5.6	5.6	4.4	4.0
K_2O	0.54	0.65	0.75	0.24	2.1	1.0
P_2O_5			0.2	0.1	0.3	<0.1
Rb	6.	7.				
Ba	39.	55.	260.	200.		
Sr	106.	105.	140.	89.		
Cs		0.08				
Ni		<10	46.	22.		

*Total Fe as FeO

1. East Pacific Rise (dredged). Hart (1971a)
2. East Pacific Rise (cored). Kay et al. (1970)
3. and 4. Atlantic (dredged). Aumento (1969)
5. Average icelandite from Thingmuli. Carmichael (1964)
6. Lau Basin (dredged). Hawkins (Pers. comm., 1971)

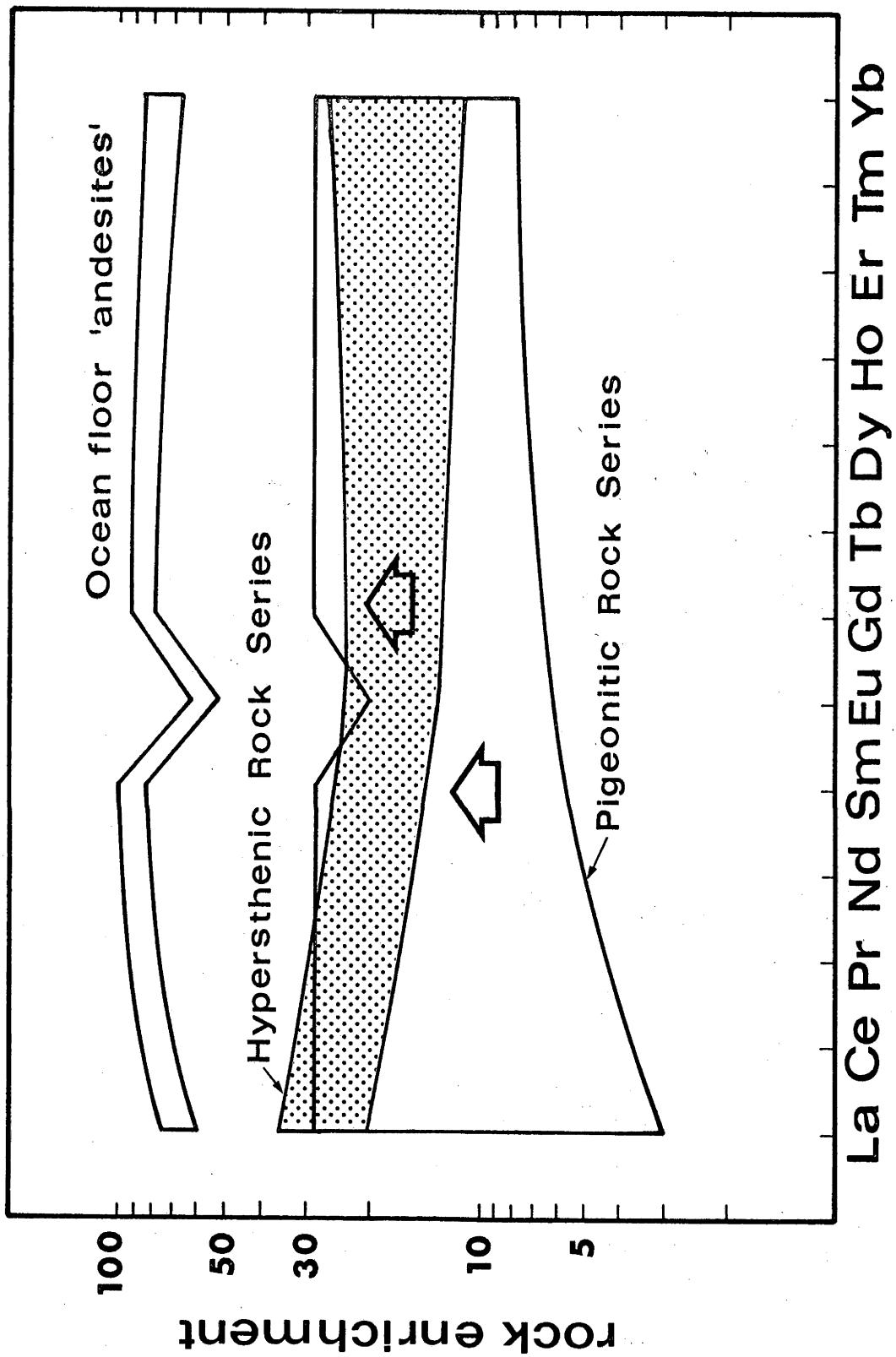

Figure 11.1 REE concentrations in intermediate rocks of Japan and the East Pacific Rise. Data sources are given in the text. Arrows indicate direction of change with increasing silica contents.

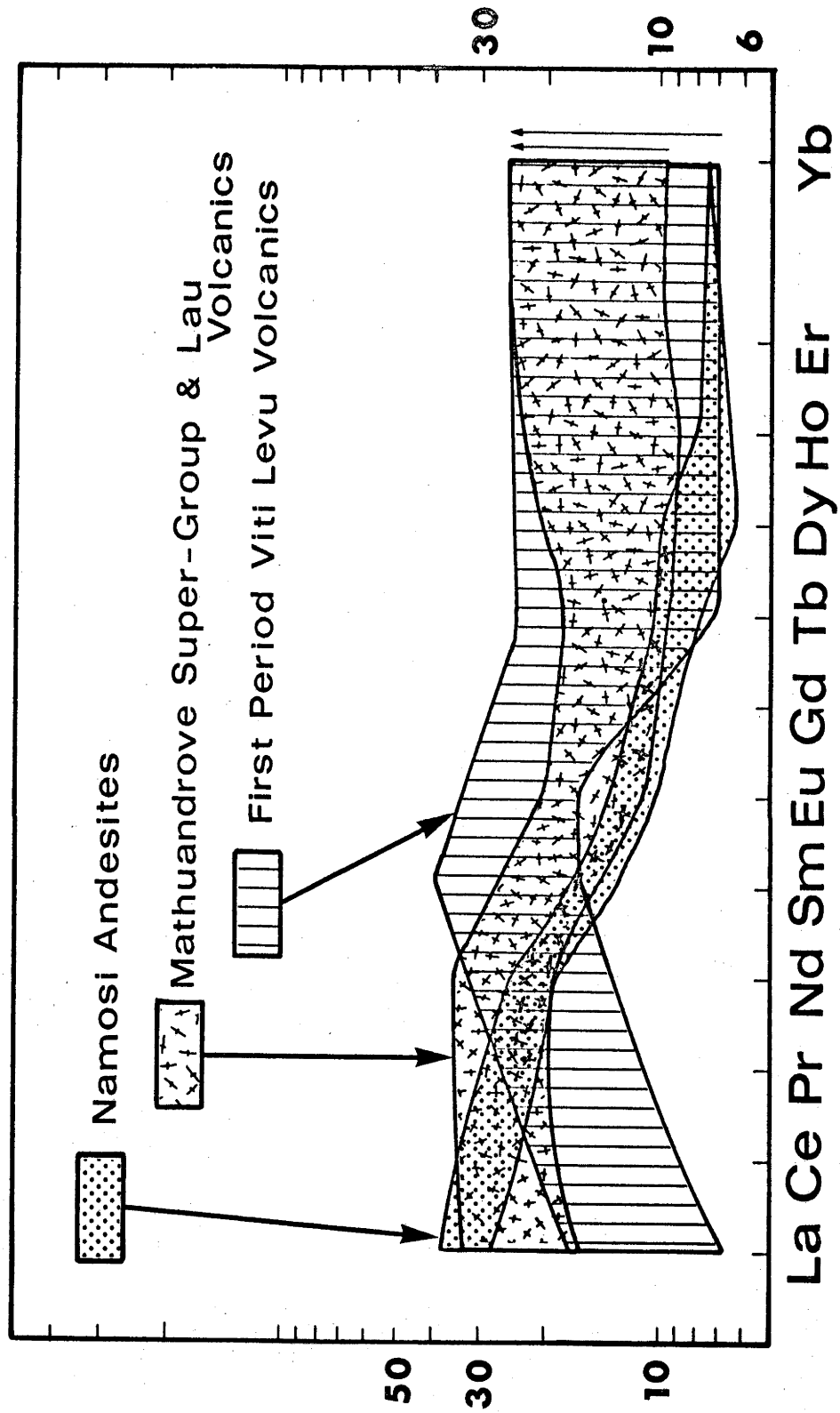

Figure 11.2 REE concentrations in calc-alkaline and tholeiitic rocks of Fiji. Data are summarized from Figures 4.5, 5.4, and 6.5.

(r_{Ti-Zr} = .77; r_{Y-Zr} = .75: both significant at the 99.75% confidence level), but in no other Fijian suite (apart from Namosi Andesites where the Ti-Zr correlation is negative) is there a significant relationship. Zr and Hf contents increase in all Fijian suites; Ti usually increases initially with Fe and then decreases. Zr/Hf ratios appear comparable or perhaps slightly higher in intermediate island arc than basic ocean floor rocks, and even higher in island arc acidic ones.

11.4 Comparison between the island arc tholeiitic series and inter-arc basin volcanic rocks

Tertiary basalts were encountered during JOIDES drilling in the Mariana and Lau Basins. Analyses of basalts dredged from the latter have been reported by Sclater et al. (1972), Hawkins and Nishimori (1971), and Hart (1971b and written comm., 1971). J. Hawkins (written comm., 1971) kindly provided me with twenty basalt and two dacite analyses. These and personal observations during the 1971 Antipodes cruise in the Lau Basin form the basis for the following comments.

Inter-arc basin like mid-ocean volcanic rocks are predominantly fresh, low-alumina olivine tholeiites with minor phenocrysts of olivine and/or bytownite. There are aphyric examples of these with 11% MgO, 200 ppm Ni, and 900 ppm Cr but only 130 ppm K, 0.2 ppm Rb, 0.002 ppm Cs, 12 ppm Sr with 87/86 ratios greater than 0.704, and 1 ppm Ba. Unlike mid-ocean floor regions, qz-normative tholeiites are relatively common in the Lau Basin, contain about 1% K_2O and, at least sometimes, contain higher Fe, Ti, Al, and large cation trace element contents than do ol-normative varieties. Finally, there are dacites (63% SiO_2) which, like mid-ocean intermediate rocks, are low in Al and Ca but, unlike them, are also low in Ti and Fe. Silicic pumice covers much of the Lau Basin floor and there is uncertainty whether it and/or the dacite flows (which were dredged near the Lau Ridge) are analogues of Coral Sea drift pumice (Bryan, 1968) originating in Tonga, are vestigial rocks from an ancestral Fiji-Lau-Tonga Ridge (see chapter 8), or were erupted in situ. They differ in composition from either Fonualei or Metis Shoal dacites (both of which have more Ca and less Na: Ewart and Bryan, 1972; Melson et al., 1970) and Coral Sea pumice. They also differ in composition from Lau Ridge dacites (which have more Ti, K, and P, and less Mg: chapter 6), and from mid-ocean intermediate rocks (Table 11.3).

Hawkins and Nishimori (1971) propose that similar low pressure fractionation occurs in both inter-arc basin and mid-ocean floor

tholeiitic magmas, i.e. via crystallization of bytownite and olivine, the prominant phenocrysts, in the approximate ratio of 3:1 (compare with Kay et al. 1970 and discussion in section 11.3). This may relate major element compositions of ol- and qz-normative tholeiites, but cannot account for their minor and trace element differences which require variations in the initial partial melts.

Island arc tholeiitic basalts are especially similar to inter-arc basin quartz tholeiites (Table 11.2). Both display somewhat erratic fluctuations in K, Ti, P, and Al concentrations. As when comparing island arc with mid-ocean volcanic rocks, the greatest difference is the predominance of basalt on the ocean or inter-arc basin floor and of basaltic-andesite in the arcs. As noted above, the Lau Basin dacite differs from arc tholeiitic dacites. A study of its trace element and isotope geochemistry was not able to be finished in time for inclusion in this thesis.

11.5 Genesis of the island arc tholeiitic series

A. Review of options

Because the island arc tholeiitic series is just one portion of a continuum which also includes the normal and high-K calc-alkaline series, this section and chapter 10 overlap. Jakes and I (1970, and Gill, 1970) and many others have summarized three options by which to explain the genesis of rocks forming this continuum: (1) melting of only underthrust lithosphere (see section 9.5); (2) melting of only sub-arc upper mantle (see section 9.4); and (3) various mixtures of the low-melting fraction of both regions. The first alternative involves a two-stage model for island arc magma genesis and was proposed in a series of papers in the late 1960's by T. Green and Ringwood, Taylor and White, and others. It was subsequently refined to accommodate the spatial variations emphasized in this thesis by Jakes and White (1970), Gill (1970), Fitton (1971), T. Green (1972), and is model A in Jakes and White (1972). These authors view the sequence of island arc tholeiitic to calc-alkaline to high-K calc-alkaline (or shoshonitic) series as reflecting decreasing amphibole/phlogopite ratios in the down-going slab (i.e. transition from amphibolite to eclogite facies assemblages) and decreasing degrees of partial melting. The second alternative has gained broad support in Japan through the influence of Kuno, Sugimura, and others, and has been even more widely accepted following Kushiro's experimental studies discussed in section

9.4. The third alternative envisages blending of partial melts from both slab and mantle sources through addition of alkali and water-rich siliceous magma from the slab to mantle overlying it. This was first proposed by Tatsumoto and Knight (1969) who suggested that the slab contribution decreased with depth. Jakes and I (1970) developed the idea further and it is model B in Jakes and White (1972) where it is implied that the slab contribution increases with depth. I shall now consider these three options in light of information presented and discussed in chapters 3 and 8 to 10.

B. Option 1: melting of only underthrust lithosphere

This has the joint attraction of providing a constantly replenished source for copious amounts of predominantly intermediate and acidic magmas and of satisfying requirements for an intimate relationship between volcanism and underthrusting (section 9.2). It faces many difficulties, however.

In-coming of subsolidus garnet occurs between 10 to 15 kb in basaltic compositions whether dry (Green and Ringwood, 1972, Figure 2) or under-saturated with water (Hill and Boettcher, 1970; Wyllie, 1971a, Figure 18-8). Velde et al. (1970) interpret a metamorphic assemblage in central France as indicating transition to garnet-amphibolite in basic rocks at 6.5 kb. Thus, rocks of basaltic composition within underthrust lithosphere beneath the volcanoes of all modern island arcs will probably occur in eclogite facies assemblages and will certainly contain garnet. In chapter 10 I argued that partial melting of eclogite having the composition of ocean floor basalt was unlikely to yield magmas with the minor and trace element characteristics of Namosi Andesites. More tholeiitic rocks, differing from Namosi Andesites as described in section 11.2, are even less likely to be explained in this way. It is difficult to see, for example, how partial melting of any garnet-bearing assemblage could yield rocks with the REE patterns shown in Figures 11.1 and 11.2. Neither the absolute concentrations of heavy REE and Y in these island arc rocks (relative to ocean floor basalts) nor their increase in more siliceous varieties is consistent with derivation from a garnet-bearing source. This apparently precludes derivation of magmas from lithosphere vertically beneath the volcanoes from which they erupt.

It may be possible that magmas do not rise vertically. Melting before the appearance of garnet (i.e. at depths <50 km) requires

temperatures >600°C if $P_{H2O} = P_{TOT}$ or ∿900°C if $P_{H2O} < P_{TOT}$ (Figure 9.3). Such temperatures at <50 km depths require considerable shear-strain heating along the uppermost slab surface. Fitton (1971) used isotherms estimated by Oxburgh and Turcotte (1970) to indicate that melting conditions are possible. As already discussed, this argument is circular because Oxburgh and Turcotte assumed temperatures sufficient for partial melting in their calculations. Moreover, the amount of shear-strain heating necessary for the shape of Oxburgh and Turcotte's isotherms is thought to be excessive (see Toksöz et al., 1971; Sass, 1971; and Hanks and Whitcomb, 1971). Temperatures predicted by the revised calculations of Toksöz et al. (1971, Figure 10), for example, do not cross the solidus for even the water-saturated condition in Figure 9.3 at pressures <25 kb. Melts formed at temperatures below the amphibole-out curve in Figure 9.3 will be water-saturated (Wyllie, 1971a, p. 203) so that their ascent to the surface is unlikely. Thus, even if magmas can rise at about a 25° angle to the vertical, it is not at all certain that temperatures sufficient for partial melting and magma movement can occur at depths where garnet is not an important subsolidus phase.

Nor is it certain that melts so produced will have the composition of island arc tholeiitic rocks. Neither the melting behavior and composition of relevant minerals nor trace element distribution coefficients for amphiboles and acidic magmas are sufficiently constrained to permit even the limited sort of quantitative tests summarized in sections 10.2 and 10.3. Two qualitative comments are, however, possible. Total fusion, an unlikely event requiring temperatures of at least 1000°C, would be necessary to produce basalts with compositions as summarized in Table 11.3. Initial melts, on the other hand, will be rhyodacites and, depending on amphibole and mica stabilities, may or may not concentrate large cations. It is difficult to foresee these rhyodacites having the high K/Rb ratios of the low-K island arc tholeiitic series variety (Table 11.1) if these melts co-exist with significant amounts of residual amphibole (assuming the amphibole/liquid distribution coefficients used by Gast [1968] and Griffin and Murthy [1969] for K and Rb are approximately correct).

Finally, variation of 87Sr/86Sr ratios within the island arc tholeiitic series or of Sr and Pb isotopes across island arcs is incompatible with derivation of all magmas from underthrust lithosphere unless disequilibrium melting occurs (see Gill and Compston, Appendix 3).

The combination of arguments presented above and in sections 10.6 and 10.8 strongly suggest that option 1 is unlikely.

C. Option 2: melting of only sub-arc upper mantle

This option can take one of two forms: (2a) in which sub-arc mantle is no different from any other upper mantle; or (2b) in which it receives water (but only water) from underthrust lithosphere (which is model 3 of Green and Ringwood, 1968). Both require partial melting of hydrous peridotite which is inhomogeneous in trace element and accessory mineral distribution. Depending on factors outlined in section 9.4 a range of basalt (but only basalt) compositions, including qz-normative varieties, can be obtained through either process. Both are consistent with the existence of low-Q, low-V mantle behind and perhaps beneath island arcs, the degree of field association between basalt and other rocks in island arcs (emphasized especially by Kuno), the similar major element compositions and trace element ratios of basalts from ocean floors, inter-arc basins, and island arcs (sections 11.3 and 11.4), and the similar REE patterns of island arc and ocean floor basalts (section 11.3 and Jakes and Gill, 1970). Option 2a is analogous to partial melting beneath mid-ocean ridges; option 2b to that beneath inter-arc basins. Differences between the volcanism of these areas and island arcs must be explained if either option is invoked.

There are more similarities between basalts of island arcs and inter-arc basins than between basalts of island arcs and the deep ocean floor (Table 11.2), which supports option 2b over 2a. The major difference common to both analogies is the relative abundance of intermediate rocks in the arcs. Both versions of option 2 require these intermediate rocks to be either: (i) residual liquids remaining after fractional crystallization of basaltic magmas; or (ii) basaltic liquids which assimilated the crustal low-melting fraction during ascent and/or crystallization (section 9.3). Both alternatives have long histories (compare, for example, reviews of them by Kuno, 1968a, and Green and Ringwood, 1968).

Taylor (1969) argued that the relatively low concentration of large cation trace elements in intermediate island arc rocks was incompatible with derivation through fractional crystallization of basalt. This depends, however, on the trace element composition of the basalt, the nature and amount of crystallizing phases, and the distribution coefficients. Basalts with very low concentrations of large cation

trace elements occur in island arcs, and crystallization of amphibole would moderate the concentration of these elements in residual liquids.

Taylor, Kaye, et al. (1969) stressed that the similarity between Ni contents of intermediate island arc rocks and the basalts with which they are spatially associated precludes a fractional crystallization relationship between them. This too depends on the factors itemized in the preceding paragraph. For example, my calculations in chapter 10 predicted a change of only 4 ppm Ni between basaltic-andesite and dacite compositions using $K_{Ni}^{amph} = 2.8$. The change would have been even less if I had used $K_{Ni}^{amph} = 1$, following Gast (1968) and Jakes' (1970) experimental results. In contrast, the ocean floor andesite analyzed by Kay et al., 1971, which formed by crystallization of olivine, plagioclase, and augite from basalt according to Hart (1971a; although REE data do not support this suggestion), has "<10 ppm Ni". Moreover, some basalts in island arcs have >100 ppm Ni (Table 5.1; Hedge, 1971) and their Mg/Ni ratios can be comparable to those of mid-ocean or inter-arc basin varieties. Thus, arguments based on large cation trace element and Ni concentrations cannot a priori preclude derivation of island arc andesite from basalt, although they could apply in certain situations.

There are, as noted in sections 11.3 and 11.4, differences between the composition of intermediate rocks from island arcs and those from ocean floors and inter-arc basins. The latter are poorer in Ca and Al and richer in Fe and Ti, as would be expected when fractionation is dominated by olivine, plagioclase, and clinopyroxene. This confirms the geochemical commonplace that the lower degree of iron-enrichment in island arc intermediate rocks requires special explanation (section 10.1). The comparison above suggests that if option 2 applies, then plagioclase crystallization must be less important and precipitating crystals must have higher Fe/Mg and Fe/Si ratios in island arcs. Both are consistent with crystallization under higher P_{H2O} or P_{O2} in the arcs.

Osborn (1969 and references therein) thought the lower degree of iron-enrichment due to earlier crystallization of magnetite. He noted this would occur under conditions of high and constant oxygen fugacities resulting from dissociation of water acquired from the lower crust. Osborn's argument has been theoretically and empirically criticized. Burnham (1967) emphasized that it requires crustal P_{H2O} to

exceed that of the magma, that O_2 diffuses very slowly in silicate melts, that only a few meters of magma adjacent to the contact would acquire appreciable water, and that conditions permitting even this are unlikely. Carmichael (1967a; Carmichael and Nicholls, 1967; Smith and Carmichael, 1968) found that magnetite was not an early crystallizing phase and that oxygen fugacities did not remain constant in Cascade lavas (which are even less iron-enriched than Namosi Andesites). Taylor, Kaye, et al. (1969) note that if magnetite crystallization is the sole reason for lack of iron-enrichment, it will also lead to lower V contents in residual liquids than are observed in intermediate island arc rocks. Finally, magnetite crystallization will not affect the Ca and Al differences noted above.

Qualitatively there are two other alternatives by which to explain lack of iron-enrichment. One involves high pressure crystallization of eclogite (Green and Ringwood, 1968) but is subject to the criticisms itemized in section 10.7. The second is more consistent with option 2b than 2a and involves fractionation of more hydrous basaltic magmas in island arcs than elsewhere. This will result in less crystallization of plagioclase and more of amphibole and magnetite. Water occurs as H^+ and OH^- in silicate magmas and H_2 can diffuse more easily than O_2 because of its smaller molecular radius (Burnham, 1967). Thus magma with initially high water contents may maintain relatively high though falling oxygen fugacities in a complex buffering relationship between the amounts of water, O_2, and condensed phases in the system (see Carmichael and Nicholls, 1967, pp. 4673-76, and Hamilton and Anderson, 1967, pp. 471-477). One would expect there to be a continuum between magmas of, for example, the Skaergaard, the ocean floor, inter-arc basins, and various suites in island arcs (see below). The "continuum" would be punctuated with discrete breaks as various minerals become stable or unstable. Amphiboles analyzed by Green and Ringwood (1968) from experiments at 9 to 10 kb on water-deficient basaltic compositions were both sub-silicic and had relatively high Fe/Mg ratios. Crystallization of such amphiboles, clinopyroxene, and minor magnetite instead of olivine, plagioclase, and clinopyroxene may, qualitatively, explain the differences between the composition of intermediate rocks of island arcs and ocean floors or inter-arc basins. The contrasting behavior of Ti and Zr in island arc and ocean floor volcanics also suggests greater precipitation of Fe-Ti oxides in the arcs.

This is approximately the conclusion reached in section 10.7 where it was shown to be consistent with major and trace element (including V) variations within the Namosi Andesite suite. But the Namosi Andesites represent just one part of the island arc continuum. If all island arc magmas originate in the upper mantle with no more than water being contributed by underthrust lithosphere, then all transverse variations in magma compositions across island arcs must: (1) be inherited from the source material; (2) be due to differences in minerals crystallized during cooling; and/or (3) reflect differences in interaction with the crust or mantle during ascent and crystallization. Following the reasoning outlined in sections 9.2, 9.3, and 10.8, the third process is unlikely to be the principal cause of the variations but may contribute to them. I cannot quantitatively assess these three influences so instead offer a few more qualitative comments.

Maximum iron-enrichment and average iron contents in intermediate rocks decrease with distance from the trench (chapter 3). Amphibole and biotite become more common phenocrysts in the same direction and the lowest silica content of rocks in which they appear decreases. These features suggest that P_{O2} and P_{H2O} in magmas may increase with distance from the trench across island arcs. This is consistent with the suggestion (which is inherently difficult to verify by rock analyses) that Fe_2O_3/FeO ratios in magmas likewise increase away from the trench (Jakes, 1970; section 11.2; and data for the Late Miocene volcanics of Fiji). Thus, it may be possible to explain consistent transverse variations in the geochemistry of island arc magmas by appeal to factors (1) and (2) in the preceding paragraph. That is, major variations in the geochemistry of rock suites may reflect smaller variations in the geochemistry of parental basalts upon which are superimposed effects of relatively low pressure fractional crystallization in which water and oxygen partial pressures increase with distance from the trench. (I specify relatively low pressures because, unless heavy REE and Y garnet/liquid distribution coefficients are <1 instead of ∿10 as currently thought, no known island arc rock suite is compatible with an evolution involving fractional crystallization of garnet.) If so, geochemical effects due to amphibole and magnetite crystallization should become more apparent and those due to plagioclase should become less apparent with distance from the trench.

There are three major problems with this proposal apart from its vagueness. First, plagioclase crystallization was found to be almost

as important as amphibole crystallization in the explanation of differences between members of the Namosi Andesite suite which was presented in section 10.7. Moreover, Al_2O_3 seems to begin decreasing at lower silica contents in calc-alkaline than in island arc tholeiitic suites (compare Figures 4.2, 4.4, 5.2, and 6.3). This may imply that plagioclase precipitates earlier in the cooling history of magmas further from the trench.

Second, differences in isotopic composition are difficult to explain. If these variations are primary they require a very special upper mantle as Pb and Sr must become less radiogenic with distance from the trench or with depth (Tatsumoto and Knight, 1969; Gill and Compston, Appendix 3). This in turn implies that Rb/Sr, Th/Pb, and U/Pb ratios in sub-arc mantle decrease away from the trench or with depth. However, Th/Pb and U/Pb ratios in volcanic rocks increase across modern Japan and, according to the reconstruction presented in chapter 8, increased across the Lau-Colville Ridge in the Late Miocene. This requires, at least in Japan where Pb isotopes are known, a prior and preferential extraction of Pb relative to U and Th which increased in efficiency with distance from the modern trench. Preferential Pb extraction has been postulated frequently (e.g. Oversby and Gast, 1970) and appears to be plausible. If, however, Rb/Sr ratios increase across island arcs while 87Sr/86Sr ratios decrease (see data in Hedge and Knight, 1969; Gill and Compston, Appendix 3; and Table 3.1), similar arguments must be developed which will require preferential extraction of Sr relative to Rb which is much less likely (see distribution coefficients in Griffin and Murthy, 1969, for example). Variations in Sr isotopes within some island arc tholeiitic series are similarly difficult to explain if its isotopic geochemistry is fixed by the mantle source.

Third, there is no a priori reason to expect higher water contents in magmas further from the trench. If, according to option 2b, water is inherited from slab dehydration, one might expect higher water contents closer to the trench. One must also explain why all volcanoes occur above earthquake foci deeper than those likely to be associated with the amphibolite to eclogite boundary within the slab (section 9.5) and deeper than the 40 km limit for generation of qz-normative basalts from water-saturated peridotite (section 9.4). The difference in magma water contents may reflect degree of fusion. If slab dehydration and island arc volcanism are in any way related, the position of volcanoes requires movement of mantle diapirs (as envisaged by Green and Ringwood,

1967) but in a direction away from the trench as well as towards the surface. This was predicted by McKenzie's (1969, pp. 12-19) calculations of flow lines for mantle underthrust by lithosphere.

It is hard to tell whether differences inherited from a basalt parent and modified by crystallization under varying water and oxygen partial pressures can account for transverse variations in potassium and other large cation abundances and ratios or for differences between the three REE fields defined in section 11.2 for island arc rocks. The variations discussed throughout this thesis require lower concentrations of large cation and light rare earth trace elements in basalts generated closer to the trench, and higher ratios of these elements relative to silica in phases crystallizing from those basalts. The different shape of REE patterns may be inherited from the basalt parent (Masuda, 1967, 1968). The greater enrichment of heavy REE and Y and the greater depletion of Eu in siliceous members of more tholeiitic series may reflect more crystallization of plagioclase and less of amphibole. Amphibole can have heavy REE and Y partition coefficients greater than unity in silica-rich magmas (Nagasawa and Schnetzler, 1971) whereas pyroxenes and plagioclase apparently cannot. (For example, had I used Nagasawa and Schnetzler's coefficients in the calculations summarized in Table 10.6, heavy REE and Y contents would have been predicted to decrease rather than increase.) Finally, Rb/Sr ratios might be expected to be higher relative to silica contents as a result of plagioclase rather than amphibole-dominated fractional crystallization. Variability within Rb and Sr data from various Fijian suites and within data for suites from Tonga (Ewart, pers. comm., 1971) or the South Sandwich Islands (Baker, pers. comm., 1971) is too great to identify any differences. Data summarized in Table 3.1 are at least consistent with the proposition that Rb/Sr ratios at a given silica content increase with distance from the trench.

Thus a version of option 2b involving fractionation of basalts at relatively shallow depths and under varying water and oxygen partial pressures, though problematical and vague as developed here, appears to be consistent with more aspects of island arc magma geochemistry than either options 1 or 2a. Effects of crustal or mantle wall rock reactions are difficult to assess but seem less important than differences in the parental magma composition and the amount and nature of minerals crystallizing.

D. Option 3: mixing of the low-melting fractions of underthrust lithosphere and sub-arc mantle

This option envisages that a variety of melts ranging from water-rich rhyodacites (derived from amphibolites) to water-poor andesites (derived from eclogites) originate during partial fusion of the down-going slab, rise into hotter overlying mantle, and cause melting there. The version involving rhyodacite is a variation of option 2b in which hydrous phases within the slab decompose at rather than below the solidus, and is more likely to cause melting in overlying mantle than the latter case. Both are difficult to assess quantitatively but either might avoid some of the problems encountered by option 2b.

First, option 3 helps to explain the apparent limitation of abundant volcanic rocks with intermediate compositions to areas underthrust by lithosphere without requiring the composition of such rocks to be consistent with primitive partial melts (or fractionation products thereof) of that lithosphere.

Second, it may resolve the dilemma of isotopic variations within and between island arc suites by providing an additional source for Pb and Sr. This was pointed out by Tatsumoto and Knight (1969) and Jakes and myself (1970). If Sr and Pb isotopes in island arc magmas are not affected by the crust through which they pass, then the relative proportions of slab and mantle contributions to these magmas can be deduced if one knows the isotopic composition of the slab. As usual, there are too few constraints. One cannot predict whether Pb and Sr in underthrust lithosphere will be more or less radiogenic than in the overlying upper mantle (unless oceanic sediments are involved in which case Pb and Sr will be more radiogenic in the slab). If the slab is more radiogenic, then its contribution decreases with depth across island arcs and is less in rocks with lower silica contents and Rb/Sr ratios in some island arc tholeiitic series. The latter especially seems more consistent with addition of alkali and silica-rich magma from the slab than the converse which is required if the slab is less radiogenic than overlying mantle. This is similar to the conclusion of Tatsumoto and Knight (1969) and Tatsumoto (1969) but their attempts at quantification, based on the isotopic composition and concentration of Pb and requiring underthrusting of sediments, are just as unable to explain the isotopic composition and abundance of Sr in island arcs as the cases discussed in section 10.4. The difference in behavior of Sr and Pb isotopes in Tongan volcanic rocks (the former increase according

to Gill and Compston, Appendix 3; the latter are constant according to Oversby and Ewart, pers. comm., 1972) is difficult to reconcile with any of the options under discussion.

The option 3 magma mixing process, coupled with inherent heterogeneities within both slab and overlying mantle, the lingering possibility of crustal or mantle wall rock reactions, and local differences in water and oxygen partial pressures, leads to endless permutations and variations of magma compositions. This is consistent with the diverse geochemistry of island arc volcanics as, for example, in Vanua Levu, Fiji (chapter 5). How this process will influence the transverse variations in magma composition across island arcs is difficult to discuss even qualitatively. It certainly implies that the relationship between these geochemical variations and the underthrusting of lithosphere is real and causal but nevertheless indirect and complex.

Chapter 12. The Origin of Shoshonites and Other Magmas in Third Arcs

12.1 Introduction

If speculations in chapters 2 and 8 are approximately correct, Fiji began about 5 to 7 m.y. ago to be an area in which magmas could not originate through processes related to underthrusting. Plate relationships in the Fiji region apparently changed late in the Miocene and the site of Pacific underthrusting moved east at about 5 cm/yr (see Karig, 1970, 1971b) leaving the Lau Basin in its wake. West of Fiji a different but related adjustment may have occurred, perhaps beginning sooner, involving fracturing of a once-continuous New Hebrides-Fiji-Lau Ridge and clockwise rotation of the New Hebrides portion more or less along the trace of the Hunter Fracture Zone (see Gill and Gorton, Appendix 3). This may be responsible for the back-bent shape of the northern Lau-Colville Ridge where the principal Fiji islands and post-Miocene volcanics occur.

However such plate rearrangements occur, they would obviously influence magmatism in Fiji. As discussed in chapters 4 to 8 (note requirements itemized in section 8.3), pre-Pliocene volcanism was predominantly intermediate but thereafter it became mostly basic and included absarokite-shoshonite-banakite suites (Sabeto and Tavua), high-Ti alkali basalts (at Taveuni and Koro), and tholeiitic basalts (at Nakobalevu), as well as suites transitional between shoshonites and alkali basalts (at Rakiraki and some Lomaiviti Islands) and between alkali and tholeiitic basalts (at Bua, Korobasaga, and Nairai) having varying degrees of silica-saturation and varying concentrations of Ti, K, and trace elements. Their over-lapping diversity mocks attempts, as above, to subdivide and name.

As the Lau-Colville Ridge became a third arc (for definition and etymology see Karig, 1970) it presumably became less underthrust, or at least more deeply underthrust by lithosphere, and its tectonic position gradually changed from one analogous to that of modern Tonga to its own present situation. This would mean (a) removal of a source of water and (if option 3 of chapter 11 applies) siliceous magma as discussed in sections 9.5 and 11.5, and (b) lowering of temperatures. Both depend critically on the mechanism by which subductive foci migrate and on temperature time-constants. Oxburgh and Turcotte (1971) suggest that 10 to 20 m.y. would pass before thermal equilibrium is

re-established. The seismic normality of upper mantle beneath the modern Lau-Colville Ridge noted in section 9.4 suggests this may be an overestimate (if my reconstruction of events is correct), but one cannot be confident without heat flow measurements from Fiji. Decreasing temperature and availability of water or water-rich siliceous magma would lead to gradual cessation of volcanism, a more basic character of that which did occur, smaller degrees of melting and amounts of magma, smaller amounts of hydrous phases in the upper mantle beneath Fiji when and if its acquired water is dissipated, and enhanced propagation and less attenuation of seismic waves in that mantle. The Pliocene to Recent record of Fiji seems consistent with such an evolution, as I shall now discuss.

12.2 Shoshonitic associations

A. Geographic distribution and descriptive geochemistry

Joplin (1968 and references therein) suggested that certain suites of predominantly basic potassium-rich rocks be called "shoshonitic associations" (following Iddings' name for rocks in Wyoming where Shoshone Indians once lived). Such rocks occur at the Sabeto and Tavua centers of Viti Levu, are discussed in sections 4.4B and C, and are compared to high-K calc-alkaline rocks in section 10.8.

K_2O/Na_2O ratios are about 1.0 and TiO_2 contents about 0.6% in Fijian shoshonitic rocks, in sharp contrast to those in alkali basalts. Alkali and alkaline earth minor and trace elements are uniformly enriched in the shoshonitic association (i.e. K/Rb, K/Sr, and K/Ba ratios are similar to those of alkali basalts) but Ti, Zr, Hf, Nb, Th, U, and the light REE are not enriched. Sr isotope ratios are lower than observed in alkali basalts of comparable K/Na ratios (Peterman and Hedge, 1971).

The shoshonitic associations summarized by Joplin occur in either island arcs or continental rift valleys. To her instances of the former can be added New Guinea (Smith, 1970; MacKenzie, 1970; Ruxton, 1966), Kamchatka (Ehrlich, 1968), Puerto Rico (Jolly, 1970), and northwest Honshu, Japan (Oshima-ōshima and Ichinomegata: Ono, 1962). These locations, Fiji, and eastern Java, are all areas of complex subduction with or without proximity to inter-arc rifting, and this may apply equally to the Wyoming type-locality (Christiansen and Lipman, 1972; Lipman et al., 1971; and Scholz et al., 1971). These island arc shoshonitic associations can be distinguished from rift valley

varieties by their low Ti and related element contents. They are spatially and temporally associated with (but in no consistent way), and less voluminous than, high-K calc-alkaline volcanics in Wyoming, New Guinea, Kamchatka, and Honshu (compare Smith, 1970, Figure 12, with Peterman et al., 1970, Figure 2). Calc-alkaline volcanics are also associated with Fijian shoshonitic rocks (section 4.4E) but are volumetrically insignificant and less potassic than in these other areas. They are, however, enriched in most large cation trace elements and radiogenic Sr relative to other, older Fijian intermediate rocks. Their age relationship to the shoshonites is unclear.

In detail there are also differences between the Fijian and Wyoming shoshonitic associations themselves. There are fewer analyses for the latter but they show: higher concentrations of Pb, Th, and U; higher Th/U, similar Pb/U, and lower K/U ratios; similar K, Rb, Ba, and Sr contents and ratios: higher and more variable 87Sr/86Sr ratios; higher Ti and Zr contents and lower Ti/Zr ratios; and higher concentrations of light REE and higher La/Y ratios (Nicholls and Carmichael, 1969; Peterman et al., 1970: note differences in reported Rb and Sr contents but similarity in Rb/Sr ratios for identical samples, suggesting contrasting mass absorption estimates in XRF analyses).

B. Genesis

In this section I use "shoshonite" to denote the parental magma of the association without reference to silica content and assume it is a relatively unfractionated partial melt. Theories concerning its derivation must explain the differences between Fijian and Wyoming varieties, between shoshonites and alkali basalts, and between shoshonites and high-K calc-alkaline rocks. The first comparison suggests crustal involvement as crust is older (∿3.0 b.y.), thicker (40 to 50 km), and more fractionated in Wyoming than in Fiji (see references in Peterman et al., 1970). Peterman et al., however, concluded that bulk assimilation of crust was incompatible with the isotope geochemistry of Wyoming shoshonites. Lower thermal gradients after the Miocene would decrease the likelihood of crustal anatexis in Fiji and initial melts of lower crust (whether granodioritic, eclogitic, or amphibolitic) would have major element compositions contrasting sharply with shoshonites (chapter 9).

The simplest explanation for differences between shoshonites and alkali basalts is heterogeneous distribution of the elements, mostly

minor and trace, in which they differ. This explanation is unlikely. The regional distribution of these respective rock types and differences in the nature of suites with which they are spatially and/or temporally associated suggest tectonic rather than random geographic control. The 87Sr/86Sr ratios and REE contents of Fijian shoshonites suggest that their relative enrichment in K, Rb, Sr, and Ba does not reflect melting of more primitive upper mantle than that which is parental to alkali basalts although they are among the few terrestrial rocks with chondritic K/U ratios. Conversely, shoshonites are unlikely products of previously depleted mantle as this would require earlier melting which concentrated Ti, Zr, Hf, Nb, Th, U, and the light REE but left behind Rb, Sr, Ba, Pb, and P.

A more promising alternative is that initial and/or residual phases in the respective source regions differ. Jakes and White (1970, 1972), Jakes and Smith (1970), and Yoder and Upton (1971) suggest that shoshonites are partial melts of a mica-bearing source. Jakes, White, and Smith emphasize mica-bearing eclogite within underthrust lithosphere; Yoder and Upton mention mica-lherzolite. Green (1972) suggests, instead, an amphibole-bearing peridotitic source.

Either amphibole or mica will occur in hydrous upper mantle at depths limited by total pressure and in amounts limited by alkali and water contents. Green (1972) estimated that pyrolite can contain about 30% low-potassium amphibole when saturated with water; mica contents would be considerably less. Small degrees of non-modal melting without residual mica are necessary if the geochemical similarities between shoshonites and phlogopites reflect derivation from a mica-bearing source (compare results and assumptions of Gast, 1968; Griffin and Murthy, 1969; Shaw, 1970). This may explain shoshonite K/Na, K/Rb, K/U, and K/La ratios (see Jakes and White, 1970, plus data from Kleeman, 1970, Higuchi and Nagasawa, 1969, and Schnetzler and Philpotts, 1970). It does not, however, explain the low Ti, Zr, Hf, and Nb contents (compare shoshonite analyses with inferred results of mica melting in Figure 8 of Flower, 1971) nor the high Sr contents and low K/Sr, Rb/Sr, and Ba/Sr ratios of shoshonites.

Alternatively, residual aluminous amphiboles may buffer shoshonite magma compositions. Titanium-rich amphiboles (kaersutites) occur in ultramafic nodules and as near-liquidus phases in basalt compositions containing a few per cent water. Their significance and geochemistry

has been summarized by Kesson and Price (1972) and Mason (1968). They have 4 to 7% TiO_2, 1 to 2% K_2O, and Na_2O/K_2O ratios $\sim$1.5 so that melts in equilibrium with significant residual kaersutite could have low Ti contents and Na/K ratios. There are, as usual, problems. Most kaersutites have REE patterns similar to that of absarokite 895 (Figure 4.5), but melts in equilibrium with them will be enriched in light REE according to the distribution coefficients measured by Schnetzler and Philpotts (1970) or Higuchi and Nagasawa (1969). The amphibole analyzed by Varne (1970) and Varne and Graham (1971) is even less likely to dominate a shoshonite source region in terms of Ti and REE contents or Na/K and La/Yb ratios. Kleeman (1970) found about 0.5 ppm U and Kesson and Price (1972) found <1 ppm U in kaersutites, which is slightly in excess of that in coexisting clinopyroxenes, but is ambiguous in relationship to shoshonite genesis. Kesson and Price inferred kaersutite/liquid distribution coefficients less than and equal to unity for Zr and Nb, respectively, making kaersutite an inefficient "sink" for these elements.

K, Rb, Sr, and Ba relationships are also ambiguous because of uncertainties about initial concentrations and ratios, the possibility of co-existing amphibole and mica, and absolute distribution coefficients. If increased amounts of amphibole in upper mantle once underthrust by lithosphere reflect increased water contents due to slab dehydration or partial melting, then that upper mantle may have K, Rb, Sr, and Ba concentrations in excess of, and ratios different from those in other upper mantle. This frustrates attempts to apply the methods used by Griffin and Murthy (1969) or Gast (1968, Figure 10).

If shoshonites originate within upper mantle which is unenriched in K, Rb, Sr, and Ba, then shoshonites represent 1 to 2% partial melts if these elements were initially present in the concentrations suggested by Gast, Griffin and Murthy, and Armstrong (1968), and then were totally dissolved in the magma. If magma is in equilibrium with amphibole or mica, it must represent less than 1% melting. Such melts will probably be much more highly undersaturated than absarokites (Bultitude and Green, 1971).

I can think of no explanation compatible with all observations. The behavior of accessory minerals is important for minor and trace elements (see section 10.5) especially when the degree of melting is small. Refractory minerals in the shoshonite source region must

contain U and Th but not P; Zr and Nb as well as Ti; and REE, especially light REE. These requirements suggest perovskite group minerals (see Vlasov, 1966, II, pp. 416-429, and Carmichael, 1967b) which occur at mantle pressures (Ringwood, 1970) and are stabilized relative to diopside plus ilmenite by high oxygen pressures and low silica activities (Verhoogen, 1962).

It appears even more difficult to explain the genesis of shoshonites than other island arc volcanic rocks. Differences between shoshonites and alkali basalts probably reflect higher water contents and correspondingly different refractory phases in shoshonite source regions. Differences between shoshonites and high-K calc-alkaline rocks in Fiji may indicate a component of partial melt from underthrust lithosphere in the latter. Fijian high-K calc-alkaline rocks cannot be differentiates of Fijian shoshonites (compare samples 390 and 894). Differences between Fijian and Wyoming shoshonites may indicate crustal involvement in the latter or in both.

C. Evolution of the shoshonitic association

Because most rocks from the Tavua and Sabeto centers are strongly porphyritic it is unclear whether or not they reflect a liquid line of descent. Eruptions at the Tavua center were predominantly explosive. The most common lava type in both areas is absarokite (<52% SiO_2). At Tavua the majority of these contain olivine and clinopyroxene phenocrysts (usually 5 to 10 and 20 to 40 modal per cent, respectively), have $Mg/Mg+Fe^{+2}$ ratios >0.65, and <13% Al_2O_3; I shall call them high-Mg absarokites. Other Tavua and most Sabeto rocks (including aphyric sample 898) have lower MgO contents and $Mg/Mg+Fe^{+2}$ ratios, >16% Al_2O_3, and are geochemically and mineralogically very similar to one another (section 4.4C); I shall call these low-Mg absarokites. In principle, either could be a "parental magma": the former giving rise to the latter through fractional crystallization (as suggested by the predominance of high-Mg absarokite at Tavua and the time-sequence of rocks there); or the former representing crystal accumulation in the latter (as suggested by the predominance of low-Mg absarokites at Sabeto and their similarity in composition to monzonites from plugs thought to represent parental volcanic centers: Dickinson, 1968b). If high $Mg/Mg+Fe^{+2}$ ratios are not effects of crystal accumulation, then high-Mg absarokites can be relatively unfractionated mantle-derived melts; otherwise neither variety of absarokite can.

Evaluation of these possibilities requires quantitative assessment of the effects of crystal addition or subtraction. This is hampered by lack of quantitative information about relevant mineral compositions, by variability of minor and trace element concentrations, and by the ubiquity of low-temperature alteration in the form of zeolites and calcite (Appendix 2; Crook, 1963). It appears, however, that the following geochemical behavior accompanies the transition from high-Mg absarokite to low-Mg absarokite to shoshonite in Fiji: Fe is not enriched whereas Ca and Al are; Na, K, P, Rb, Ba, Sr, Pb, Th, U, light and heavy REE, Zr, Sn, and Hf become enriched rather uniformly; $^{87}Sr/^{86}Sr$ ratios remain constant; Ti may be slightly more abundant in low- than high-Mg absarokites but Ti/Zr ratios are definitely lower in the former. K/Rb and Ba/Rb ratios tend to be lower and Rb/Sr and K/Sr ratios higher in shoshonites than absarokites and in banakites than shoshonites, although there are exceptions.

It may be possible to explain the major element variations by adding or subtracting the principal phenocrystic minerals (olivine, augite, and magnetite), although computer attempts to do so were unsuccessful. Lack of success may reflect uncertainties about pyroxene compositions (see section 4.4C) but was encountered when using either compositions consistent with Ibbotson's (1967) or Dickinson et al.'s (1968) refractive index measurements. If the latter and more likely pyroxene compositions represent those of Fijian shoshonite association phenocrysts, crystallization of several per cent magnetite together with olivine and that augite will be necessary to prevent iron-enrichment (a feature common to most rock suites characterized by high K/Na ratios: Coombs and Wilkinson, 1969; Joplin, 1968). Fractional crystallization of these three minerals could explain most of the minor and trace element characteristics noted above except the tendency for K/Rb and Ba/Rb ratios to decrease. Fractional crystallization involving feldspar or a feldspathoid is not compatible with these characteristics.

The explosiveness of Tavua eruptions, ubiquity of zeolitization, frequent occurrence of biotite, and absence of evidence for plagioclase crystallization during magma evolution, suggest that shoshonitic magmas were relatively water-rich. Liquidus geometry requires crystallization and decreased solubility of volatiles in such liquids upon pressure release which would explain the abundance of phenocrysts, including

plagioclase, and the explosiveness of eruptions. By inference from experimental work on alkali basalt (Essene et al., 1970) and mugearite compositions (Irving, 1972), mica and amphibole will probably crystallize from high-Mg absarokite containing a few per cent water, joining olivine ± clinopyroxene at some temperature below but near the liquidus. Fractional crystallization involving mica is incompatible with the geochemistry of Fijian shoshonitic rocks, but variations in concentrations and ratios of K, Rb, Sr, and Ba appear consistent with olivine + clinopyroxene + minor kaersutite crystallization.

In rocks of the Wyoming shoshonite association analyzed by Nicholls and Carmichael (1969), K, Rb, Ba, and REE contents decrease and K/Rb and Ba/Rb ratios increase in the sequence absarokite to shoshonite to banakite. Sr 87/86 ratios vary erratically (Peterman et al., 1970). These differences may indicate fundamentally different evolutionary trends or merely insufficient sampling of the Wyoming suite. Ferromagnesian trace elements are abnormally high in Nicholls and Carmichael's shoshonites and banakite and K_2O concentrations in their single absarokite and banakite samples differ considerably from the averages reported for such rocks by Iddings (1899).

As noted in section 4.4C, Sn and Sb concentrations increase in the sequence absarokite to shoshonite to banakite at Tavua, Te contents are erratic, and Au is unevenly distributed within individual samples. Cu and K contents are positively correlated in analyses of Sabeto samples ($r = .80$; significant at 99.5% confidence level) but not in those from Tavua where Cu contents are lower in shoshonites than absarokites and in the banakite than shoshonites. Thus, the solubility of Au and chalcophile element ions and complexes in alkali and water-rich magmas appears complicated but relevant to Au-Ag telluride mineralization at Tavua and Cu sulfide mineralization in the Sabeto area.

12.3 Geochemical significance of Pliocene to Recent basalts

Post-Miocene basic volcanics of Fiji, of which shoshonites are just one kind (see summary in Table 8.2), are among several Melanesian suites which show the limits of the "andesite line" concept. The Fijian rocks are analogous in composition to volcanics of comparable age in the New Hebrides (M. Gorton, pers. comm.; Colley and Warden, in prep.), New Georgia (Stanton and Bell, 1969), the New Ireland and Bismarck Archipelago regions (R.W. Johnson, pers. comm.), and the Torres Strait Islands, Papua (Wilmott, 1972). Together these define a broad continuum

between shoshonitic, alkali, and tholeiitic basalts. I have too little information concerning the age and composition of the Fijian examples to warrant extended discussion, but it will be important to assess regional and temporal variations in their geochemistry because they may have monitored thermal, mineralogical, and compositional changes in the upper mantle during migration of a subduction zone. As usual there are too few constraints and too much variability to draw firm conclusions at the present time, but the following summary may provide guidelines for further study.

The major element compositions of post-Miocene Fijian basalts are not unusual and there seems to be no consistent variation in the degree of silica-saturation with time, although the youngest basalts (2000 to 700 years old) are ne-normative and some contain ultramafic nodules. Similarly, there are few consistent variations among minor and trace elements. Na/K ratios seem generally to increase at a given level of silica-saturation and the most recent basalts (at Bua, Taveuni, and Koro) have the highest Ti contents. Rocks from these three centers and from YanuYanu islet (Vanua Balavu, Lau) share the following compositional features: 1.8 to 3% TiO_2; Na_2O/K_2O ratios >3; 2 to 3 ppm Th; 0.5 to 1.0 ppm U; Th/U and K/U ratios of 2.5 to 4.5 and 9-16,000, respectively; ∿30 ppm Y; and >150 ppm Zr. All of these element abundances exceed those of other Fijian basalts, and the Y, Th, U, and K contents and ratios are similar to those found in Hawaiian alkali basalts (Heier et al., 1964). Ti and Zr contents are lower than those in Hawaiian or Samoan alkali basalts (Hubbard, 1969, 1971) but Ti/Zr ratios are similar (70 to 80).

Sr isotope ratios vary from 0.7032 to 0.7041. Only one sample from most sites was analyzed so that representativity is questionable. If they are representative, however, there are no apparent correlations between 87Sr/86Sr ratios and age of volcanism or K/Na+K ratios (Peterman and Hedge, 1971) or any other element abundances or ratios.

These variations suggest magma derivation from geochemically heterogeneous upper mantle, the mineralogy (and therefore physical properties) of which gradually became less like that of anomalous sub-arc mantle and more like that of normal sub-oceanic mantle. That is, upper mantle mineralogy changed from that which yielded shoshonite to that which yielded alkali basalt upon small degrees of partial melting. This might indicate decreased stability of kaersutitic

amphibole with time. It will be difficult to distinguish differences due to time-dependent variations from those due to source heterogeneities or depth and degree of melting. But as more data become available, especially precise radiometric ages and trace element and isotope analyses, it may prove instructive to compare the evolution of Neogene volcanism in the western USA (Christiansen and Lipman, 1972; Scott et al., 1971) with that of Pliocene to Recent volcanism in Fiji and Pleistocene to Recent volcanism in the New Hebrides.

Of these suites I was able to study only the one from Taveuni in any detail. The major element compositions of my samples, which probably represent only the most recent volcanic episode, can be related to one another by adding or subtracting <10% olivine ($\sim$Fo85, which is appropriate for these magma compositions: Irving, 1972, Figure 2.6) and minor clinopyroxene ($Mg_{41}Ca_{43}Fe_{16}$: section 7.1), but not plagioclase. This is qualitatively consistent with the minor and trace element geochemistry. Mg is significantly correlated with Ni, Co, and Cr but not Ca, Sc, or V. K/Rb ratios increase slightly, consistent with olivine and clinopyroxene crystallization, but vary by only $\pm$11% within the suite. Na/K ratios increase from 3 to 4 which is more difficult to explain.

Chapter 13. The Geochemical Evolution of Fiji: implications of a young continent

13.1 Introduction

Even in the earliest geological studies of Fiji by Gräffe, Kleinschmidt, Meinicke, and Wichmann between 1860 and 1880 it was inferred that at least Viti Levu had continental affinities on the basis of lithologies which were encountered there. These and subsequent investigators such as Woolnough (1903), Suess (1909), Brock (1924), Davis (1927), and Ladd (1934), regarded Fiji as a relatively old, vestigial continental fragment. They disagreed on its maximum possible age and on the question of its former relationship to land masses further west. Beloussov (1968) and van Bemmelen (1966) likewise treat areas such as Fiji as vestigial continents surrounded by "oceanized" crust.

There is no positive evidence for the presence of pre-Tertiary rocks in Fiji (section 4.2A) or of a foundered Melanesian Continent in the surrounding ocean floor (section 2.2). Aronson and Tilton (1971) looked unsuccessfully in Fiji for detrital zircons which might be as old as those they found in New Caledonia and New Zealand. The evolutionary scheme developed in chapters 4 to 8 suggests that, apart from the pre-Tertiary oceanic crust on which it presumably grew, Fiji is entirely Tertiary in age unless underlain by some Mesozoic crust inherited from the Norfolk Ridge or Lord Howe Rise, or by oceanic sediments underplated and/or accreted during subduction. If this is true, Fiji is an incipient rather than vestigial continent (as was first proposed by Chamberlin in 1935) which evolved principally through processes related to island arc development, as suggested by Dickinson (1967). The role of island arc volcanism in the geochemical evolution of continents has been discussed by many authors including Taylor (1967), Markhinin (1968), and Jakes and White (1971).

Before summarizing some of the conclusions and speculations of the preceding chapters and discussing their implications for continental evolution, it may be useful to re-examine what is meant by "continental crust". Most simply, the noun "continent" denotes a large, subaerial land mass and "continental" crust is, therefore, that which underlies a continent. The problem is only one of scale. Alternatively and more relevantly, "continental" may signify any crust (wherever it occurs) having a thickness, seismic velocity profile, age, or composition

similar to that which underlies the major continents. Fiji is not a continent but its crust appears similar to that of some continental areas in thickness and seismic velocity structure. The significance of its Tertiary age and its possible composition are discussed in the following section.

13.2 The geochemical evolution of Fiji

Data recording the evolution of Fiji are not now and probably never will be adequate for a definitive description of the process. Speculation is inherent. It is for the reader to decide whether my speculations are sufficiently moored to empirically verifiable observations to make these comments valuable. The key points as I see them are the following.

(1) The crust of the Lau-Colville Ridge is 15 to 30 or more km thick. Its upper half transmits P waves at 5 to 6 km/sec, its lower half transmits them at 6.6 to 7.0 km/sec, and there may be a masked $7^1/_2$ km/sec layer beneath (Robertson, 1965, 1967; Sutton et al., 1970; Shor et al., 1971; J. Worthington, pers. comm., 1971). This does not uniquely specify the bulk composition or mineralogy of the lower crust but suggests the presence of amphibole, quartz, and/or alkali feldspar rather than of seismically faster assemblages in which pyroxene and garnet are predominant. The Lau-Colville Ridge is surrounded by basins floored with relatively thin, young, and hot oceanic crust which probably are Tertiary in age (section 2.2).

(2) The majority of the Lau-Colville Ridge crust is also of Tertiary age, subject to the qualifications mentioned in the preceding section and in section 4.2A. The evolutionary scheme summarized in chapter 8 suggests development in three stages. The first witnessed Fiji's inception as a submarine island arc ridge which grew throughout the Early Tertiary in response to plate readjustments occassioned by the northward drift of the Australian relative to the Pacific plate. By Late Miocene time this ridge had become a mature island arc which was underthrust along its northern and eastern edges by the Pacific plate. The underthrusting was accompanied by tholeiitic, calc-alkaline, and high-K calc-alkaline volcanism at increasing distances from the trench. This was the second stage. During the third and final period, which began about 7 to 5 m.y. ago, the Lau-Colville Ridge gradually became a non-volcanic "third arc". Volcanism during this transitional period diminished, was predominantly mafic, and included shoshonitic

associations as well as alkali and tholeiitic basalts and suites transitional between each of these.

(3) The chemical contribution of arc volcanism to the bulk composition of Fiji can be calculated from data in chapters 4 to 7 if the relative amounts of all volcanic products from the first two stages of activity are known. In practice, any such attempt is a subjective though educated guess. Nevertheless, using the relative abundance of modern arc volcanic products in Japan and in the Kuriles (Sugimura et al., 1963; Sugimura, 1968; Markhinin, 1968; Jakes and White, 1971) and informal estimates of relative volumes of rocks in Fiji, I calculated "average" compositions of the first and second period Viti Levu, Mathuandrove Super-Group, Lau (east and west), and Kadavu Group volcanic rock suites. These are given in Table 13.1. They are crude estimates but qualitatively instructive. By combining them in the proportion 70% first period Viti Levu + 12% Mathuandrove Super-Group + 12% Lau Volcanics (6% east and 6% west) + 3% second period Viti Levu + 3% Kadavu Group, I guessed an average composition attributable to island arc volcanism in Fiji. This is column 8 in Table 13.1. It should qualitatively indicate the composition of crust formed through processes related to island arc development unless major geochemical differentiation occurs during the rapid production of immature sediments in island arc environments. That this is unlikely is suggested by Rogers and Donnelly (1966) and by the average composition of two first period Viti Levu sediments given in column 2 of Table 13.1. The principal geochemical difference is the $^{87}Sr/^{86}Sr$ ratio which is $\sim$0.706 in these two sediments and $\sim$0.704 in the volcanics.

(4) Upper mantle rather than underthrust lithosphere was the principal source of Fijian magmas during all three stages of volcanic activity according to the arguments and speculations presented in chapters 9 to 12. During the first two stages some unspecifiable amount of water-, alkali-, and silica-rich magma probably was derived from the slab. Transverse variations in the geochemistry of island arc magmas are thought due to: (1) the amount of slab contribution (which probably decreases with depth); (2) the mantle source composition and mineralogy; (3) the depth and degree of melting in the upper mantle; and (4) differences in minerals precipitating from magmas in which water and oxygen activities increase with distance from the trench.

Table 13.1. Estimated average compositions of some Fijian rocks.

	1. First Viti Volcanic	2. Period Levu Sediment	3. Second Period Viti Levu	4. Mathuandrove Super-Group	5. Lau East	6. Lau West	7. Kadavu	8. Average Composition of Fiji
SiO_2	56.8	62.0	58.4	56.6	58.0	58.1	60.6	57.1
TiO_2	1.0	0.8	0.7	0.8	0.9	0.8	0.6	0.9
Al_2O_3	17.2	13.7	17.4	17.3	17.1	17.8	17.1	17.2
Fe_2O_3*	9.5	7.9	7.0	8.5	8.4	7.6	5.9	9.0
MnO	0.20	0.2	0.15	0.16	0.17	0.17	0.11	0.2
MgO	3.4	3.1	3.4	4.2	3.2	3.1	2.2	3.4
CaO	7.2	8.1	7.5	8.7	7.8	7.5	5.7	7.4
Na_2O	3.9	3.2	3.9	2.9	3.2	3.4	4.5	3.7
K_2O	0.55	0.78	1.23	0.69	0.93	1.28	2.9	0.72
P_2O_5	0.2	0.2	0.2	0.2	0.2	0.3	0.3	0.2
Rb	5.	8.	21.	11.	10.	20.	52.	9.
Ba	114.	100.	390.	230.	170.	200.	850.	170.
Sr	208.	270.	540.	370.	310.	350.	1350.	290.
Pb	2.	3.	4.	2.	2.	2.	8.	2.
Th	0.4	0.1	1.7	0.5	0.5	0.7	4.3	0.6
U	0.2	0.1	0.6	0.2	0.3	0.4	1.7	0.3
Y	30.	33.	22.	33.	45.	36.	17.	31.
Ni	5.	5.	8.	28.	6.	12.	6.	8.
Co	21.	10.	19.	24.	19.	16.	18.	21.
Cr	6.	7.	19.	69.	11.	30.	8.	19.
Sc	31.	31.	18.	25.	26.	19.	12.	28.
V	190.	115.	192.	209.	190.	131.	168.	190.
Cu	65.	60.	78.	73.	32.	30.	60.	62.
Zr	65.	80.	99.	67.	100.	108.	136.	73.
Hf	1.2	-	1.3	1.3	1.7	1.7	2.3	1.3

K/Rb	660.
Rb/Sr	.031
Th/U	2.0
K/U	20,000.
HPU	0.5
Zr/Hf	56.

*Total Fe as Fe_2O_3

(5) Crustal anatexis results in silicic magmas but does not necessarily lead to enrichment of large cation trace elements if Tholo Plutonic quartz-tonalites or the ignimbritic "low-K rhyodacites of the island arc tholeiitic series" (section 11.2) result from that process (sections 4.5C and 9.3).

13.3 Implications for crust and mantle evolution

A. Island arc volcanism and the origin of continents

Several authors have suggested that continents originated and grew principally through processes related to island arc development (e.g. Taylor, 1967) and that this represents a gradual instead of catastrophic differentiation of the earth with respect to elements strongly concentrated in the crust. The implication of points made in the preceding section is that this is not so.

According to my estimate, Fijian crust has an intermediate overall composition which is similar to Jakes and White's (1971) estimate, using different data, for the average composition of "mature" island arcs. The differences in Si, Al, Fe, Mg, Ca, and Na contents between "average Fijian crust" and estimates for the average composition of crust beneath continents by Vinogradov (1962), Taylor (1964), Pakiser and Robinson (1966), or Ronov and Yaroshevsky (1969), depend critically on judgments about relative amounts of rock types (i.e. basalt vs. dacite, etc.) and is probably not significant. However, the much greater differences between my estimate for the abundance of K, Rb, Ba, Sr, Pb, Th, and U in Fijian crust and others' estimates for the abundance of these elements in crust beneath continents is important. It reflects my weighting of rock series (i.e. first vs. second period Viti Levu rocks, etc.) rather than rock types, and any reasonable combination of Fijian rock series will lead to the same conclusion. Crust formed through island arc processes such as those which occurred in Fiji will have lower concentrations of large cation trace elements than are observed in rocks of continental shield areas.

There are strong arguments that crust beneath the continents is vertically zoned through processes of anatexis and dehydration, leading to upward enrichment in compositions represented by the Qz-Ab-Or ternary minimum. Heier and co-workers (Heier, 1965; Lambert and Heier, 1967, 1968; Heier and Thoresen, 1971) and Lachenbruch (1968, 1971) have likewise suggested significant upward enrichments of Cs, Rb, Th, U, and to a lesser extent K. The youthfulness of Fiji and

the relatively slow P wave velocities in its lower crust could suggest that it represents crust in which dehydration and anatexis have not yet occurred on a major scale. If so, the differences in composition noted above may only reflect bias toward enriched upper crustal rocks in estimated compositions for crust beneath the continents, and bias away from such rocks in the Fijian estimate.

There are three notable problems with this explanation which lead me to conclude that island arc volcanism of the sort recorded in Fiji, while capable of producing intermediate crust which is "continental" in thickness and seismic velocity structure, cannot by itself yield crust with the composition of that which underlies the continents. First, even if Fijian-type crust were 50 km thick and upward concentrations of Rb, Th, U, and Pb were complete (leaving the lower portion devoid of these elements), less than 5 km of upper crustal rocks with 120 ppm Rb, 11 ppm Th, 3.5 ppm U, or 15 ppm Pb (averages from Wedepohl, 1968) could form. This is geologically unreasonable and cannot satisfy crustal heat flow requirements (Heier, 1965).

Second, there is no certainty that this upward concentration is as effective as claimed by Heier and Lachenbruch. Recent work in the Australian Precambrian shield (Gray, 1971; G. Bradley, pers. comm., 1972) has found little evidence of Th, Rb, or K depletion in many high-pressure granulite facies terrains; only U seems to be systematically depleted on a regional scale. Moreover, if Tholo Plutonic quartz-tonalites or the ignimbritic "low-K rhyodacites of the island arc tholeiitic series" are anatectic melts from lower Lau-Colville Ridge crust, then anatexis is not necessarily sufficient to provide upper crustal concentrations of these elements. Finally, as was noted in section 3.4 and by Jakes and White (1971), island arc crust is likely to be inherently stratified vertically with respect to these elements. Tilling et al. (1970) also emphasized differences in the heat production capability of various rock series and argued that the exponential decrease in heat production claimed by Lachenbruch might only represent differences in the relative abundances of rock series with depth. Thus, some or even much of the inferred upward concentration of K, Rb, Cs, Pb, Th, and U could be a primary feature if crust accreted at continental boundaries somewhat as Fiji did east of Australia. To be sure, Andean-type volcanism and plutonism probably leads to a similar result (Jakes and White, 1972) through anatexis, but

neither primary nor any superimposed secondary upward enrichment within island arc crust will be sufficient to produce the amounts of Rb, Th, U, and Pb observed in the upper crust of continental shield areas.

Third, rocks with an average Rb/Sr ratio of 0.031 (Table 13.1) would not have a present-day 87Sr/86Sr ratio of 0.71 to 0.72, as is inferred for the average of the crust beneath continents (Hurley, 1968; Armstrong, 1968; Hurley and Rand, 1969) even if that crust completely formed 4.5 b.y. ago.

These observations imply that the crust beneath the continents did not form solely through processes related to island arc development. This, in turn, may suggest that some time in the past at least the cores of continents evolved through a different process which was more efficient in fractionating large cation trace elements. Several authors have suggested massive continent formation between 3.5 to 2.5 b.y. ago with subsequent accretionary additions and/or maintenance which may involve various amounts of recycled material (Armstrong, 1968; Hurley and Rand, 1969; Ringwood, 1969; Veizer, 1971). Ringwood (1969, p. 14) comments about such a suggestion: "It is conceivable that vertical differentiation played a more important role in this early rapid phase of crust formation than lateral differentiation via sea-floor spreading". This appears consistent with the foregoing analysis.

B. Island arc volcanism and differentiation of the mantle

Ringwood (1969, 1972), Dickinson and Luth (1971), and others have related plate tectonics to the long-term differentiation of the Earth through a two-stage process. Although more crust is created at mid-ocean ridges than in island arcs, this oceanic crust is returned to the mantle and no irreversible differentiation can occur by this process alone. It must be supplemented by a second stage involving physical removal of ocean floor volcanic rocks and sediments derived therefrom at oceanic trenches, or more importantly, partial melting of lithosphere during underthrusting. If, as argued in chapters 9 to 11, sub-arc mantle rather than underthrust lithosphere is the principal source of island arc magmas, then it must also be the main site of mantle differentiation in plate tectonic models. Water and heat are the primary contributions of underthrusting, although the two-stage process will also supply an unspecified amount of some elements to the overlying mantle. From there they can directly (through magma mixing: option 3 of chapter 11) or indirectly (through amphibolitization of

peridotite and subsequent partial melting: option 2b of chapter 11) contribute to island arc crustal development. Nevertheless, this underscores the importance of mantle in the wedge-shaped region between the surface and underthrust lithosphere in the evolution of island arc crust. If this mantle is the primary magma source, its limited volume (even with McKenzie-like convective mixing) and its proximity to earlier-formed continents further limit the suitability of arc volcanism as the principal mechanism of continental evolution.

Option 3 of chapter 11 is too vague to support quantitative evaluation of mixing or recycling models in which continents are viewed as having persisted in dynamic equilibrium since their formation more than 2 b.y. ago. Veizer (1971) calculated that apparent changes in the 87Sr/86Sr ratios of sea water since the Archean are compatible either with formation of all the crust underlying the continents at 3 b.y. ago followed by about 12% recycling per 300 m.y. (i.e. about 3 km^3/yr), or with formation of only 50% of such crust 3 b.y. ago and a continuous production of juvenile crust since. My speculations seem consistent with either. However, Armstrong's (1971) argument for recycling based on "adequacy factors" applies only to formation of strictly continental crust and not to formation of the island arc variety. For example, if my estimate for the composition of the Lau-Colville Ridge or Jakes and White's (1971) estimate for average island arc crust is used, and trace element contents of weathered instead of pristine ocean floor basalt are employed, then support for Armstrong's argument disappears.

13.4 Fijians and the geochemistry of Fiji

Fiji became inhabited during the waning stages of active volcanism (section 7.1). Although most of this thesis had dealt with processes and time scales in which human activity is insignificant, there are three ways in which its ideas and data might impinge on Fijians in more than an intellectual fashion.

First, knowledge of substrate geochemistry is relevant to many studies of geographical pathology, health, and disease (e.g. Cannon and Hopps, 1971). As noted in section 5.3 there is a correspondence between my analyses of some trace element concentrations in volcanic rocks and those found in stream sediments from corresponding outcrop areas. Although the problem is complex, my data and the considerable number of stream sediment analyses by the FGSD and by mining companies may contribute to regional studies of the relationship between rock, soil, and water geochemistry and nutrition and health in Fiji.

Second, although the relationship between ore deposits, rock geochemistry, and tectonism is also difficult to understand, it is nevertheless important. My proposals concerning the geochemical evolution of Fiji might help at least to clarify some ideas and thereby assist mineral exploration. For example, the coarse correlation between Cu and heavy metals with potassium in many island arc rocks, and the similar composition of rocks from Bougainville and the Kadavu Group, suggest that Kadavu could be a favorable site for mineralization. Some issues relating to Au-Ag and Cu mineralization in Fijian shoshonite associations were touched on in sections 4.4C and 12.2C.

Third, these data may help to delimit early migration patterns of the people who first populated Fiji and nearby portions of Melanesia and Polynesia. In Table 13.2 are analyses of two stone adzes thought

Table 13.2 Analyses of stone adzes from Tonga

	To 6.20	To 6.170
SiO_2	53.30	50.36
TiO_2	1.95	2.56
Al_2O_3	17.22	16.86
Fe_2O_3*	9.87	11.31
MnO	0.22	0.21
MgO	3.40	4.44
CaO	6.16	7.78
Na_2O	5.42	4.46
K_2O	1.55	1.33
P_2O_5	0.89	0.69
Sr**	340.	340.
Ba	440.	390.
La	37.	27.
Y	52.	41.
Ni	-	-
Co	15.	26.
Cr	3.	3.
Sc	12.	19.
V	81.	172.
Cu	6.	22.
Zr	130.	160.

* Total Fe as Fe_2O_3

** all trace element analyses by emission spectrography

to have been in use in Tonga about 400 B.C. (from Tongatapu site 6: Groube, 1971). These high-Ti "mugearites" (1 to 5% hyp) rocks could not have come from Tonga and must represent trade with people from another area. On the basis of data in chapters 4 to 7, there are conceivable Fijian sources for such rocks: the Bua region of Vanua Levu; Taveuni or Koro; and YanuYanu islet off Vanua Balavu, Lau. Although no available analyses of rocks from these areas match those of the Tongan artefacts, there is no geochemical reason why such rocks could not occur. Nevertheless, their petrography (A.J.R. White, unpublished manuscript) and geochemistry seem unusual for even these Fijian sites. A more likely source is Samoa or other islands north of the "andesite line". There is also a disparity between analyses of the Tongan artefacts and those of Samoan rocks (see Macdonald, 1968; Stice, 1968; Hubbard, 1971; and references therein), but the agreement seems better. Groube (1971) argues for an early Tongan-Samoan link and these analyses are at least compatible with his case.

Fiji is much more than an incipient continent. It is also a new nation. It has a promising future as well as an interesting geological, social, and political history. I hope in some way the data, ideas, and speculations in this thesis can function as a contribution to that future.

REFERENCES CITED

Adams, C.G. (1970). A reconsideration of the East Indian letter classification of the Tertiary. Bull. Br. Mus. Nat. Hist. (Geol) 19, 88-137.

Adams, J.A.S., and Gasparini, P. (1970). Gamma-ray Spectrometry of Rocks. Elsevier Pub. Co., 295 pp.

Adams, R.D. (1962). Thickness of the Earth's crust beneath the Campbell Plateau. N.Z. Jl. Geol. Geophys. 5, 74-85.

Ahrens, L.H. and Taylor, S.R. (1961). Spectrochemical Analysis. 2nd ed. Addison-Wesley, 454 pp.

Aki, K. (1968). Seismological evidence for the existence of soft thin layers in the upper mantle under Japan. J. Geophys. Res. 73, 585-594.

Albarede, F. and Bottinga, Y. (1972). Kinetic disequilibrium in trace element partitioning between phenocrysts and host lava. Geochim. Cosmochim. Acta 36, 141-156.

*Armstrong, R. L. (1968). A model for the evolution of strontium and lead isotopes in a dynamic Earth. Rev. Geophys. 6, 175-199.

———— (1971). Isotopic and chemical constraints on models of magma genesis in volcanic arcs. Earth Planet. Sci. Letters 12, 137-142.

Aronson, J.L. (1968). Regional geochronology of New Zealand. Geochim. Cosmochim. Acta 32, 669-697.

———— and Tilton, G.R. (1971). Probable Precambrian detrital zircons in New Caledonia and southwest Pacific continental structure. Bull. Geol. Soc. Am. 82, 3449-56.

Arriens, P.A. and Compston, W. (1968). A method for isotopic ratio measurement by voltage peak switching, and its application with digital output. J. Mass Spectrometry and Ion Physics 1, 471-481.

Aumento, F. (1969). Diorites from the Mid-Atlantic Ridge at 45°N. Science 165, 1112-13.

———— (1971). Uranium content of mid-oceanic basalts. Earth Planet Sci. Letters 11, 90-94.

———— Loncarevic, B.D. and Ross, D.I. (1971). Hudson geotraverse: geology of the Mid-Atlantic Ridge at 45°N. Phil. Trans. Roy. Soc. Lond. A 268, 623-650.

Avias, J. (1953). Contribution à l'étude stratigraphique et paléontologique des formation antécretacées de la Nouvelle Calédonie Centrale. Sciences Terre 2, 1-276.

———— (1972). Major features of the New Guinea-Louisiade-New Caledonian-Norolk Arc system. In: The Western Pacific: Island Arcs, Marginal Seas, and Geochemistry (ed. P. Coleman). W. Austral. Univ. Press.

Bakun-Czubarow, N. (1971). The trace transition elements in the main minerals from eclogites. Abs. Int'l. Geochem. Congr., Moscow. II, 444-5.

Band, R.B. (1968). The geology of southern Viti Levu and Mbengga. Bull. Geol. Surv. Fiji, 15.

*Anderson, D.L., Sammis, C. and Jordan, T. (1971). Composition and evolution of the mantle and core. Science 171, 1103-12.

Banno, S. and Green, D. H. (1968). Experimental studies on eclogites: the roles of magnetite and acmite in eclogitic assemblages. Chem. Geol. 3, 21-32.

Barazangi, M. and Isacks, B. (1972). Lateral variations of seismic wave attenuation in the upper mantle above the inclined earthquake zone of the Tonga island arc: deep anomaly in the upper mantle. J. Geophys. Res. 76, 8493-8516.

Barth, T.F.W. (1956). Geology and petrology of the Pribilof Islands, Alaska. U.S. Geol. Surv. Bull. 1028-F.

Bartholomew, R.W. (1959a). Geology of the Lautoka Area, Northwest Viti Levu. Bull. Geol. Surv. Fiji, 2.

———— (1959b). Geology of Savusavu Bay West, Vanua Levu. Bull. Geol. Surv. Fiji, 5.

———— (1960). Geology of the Nandi area, Western Viti Levu. Bull. Geol. Surv. Fiji, 7.

Bateman, P.C. and Dodge, F.C. (1970). Variations of major chemical constituents across the Central Sierra Nevada Batholith. Bull. Geol. Soc. Am. 81, 409-420.

Beloussov, V.V. (1968). Some problems of development of the Earth's crust and upper mantle of oceans. In: The Crust and Upper Mantle of the Pacific Area (eds. L. Knopoff, C. Drake, and P. Hart). A.G.U. Mon. 12, 449-459.

Bender, M.L. and Shultz, C. (1969). The distribution of trace metals in cores from a traverse across the Indian Ocean. Geochim. Cosmochim. Acta 33, 292-297.

Berggren, W.A. (1969). Cenozoic chronostratigraphy, planktonic foraminiferal zonation, and the radiometric time scale. Nature 224, 1072-75.

Birch, F. (1954). Heat flow from radioactivity. In: Nuclear Geology (ed. H. Faul). John Wiley and Sons, Inc., Chapter 5.

Black, P.M. (1970). Observations on White Island volcano, New Zealand. Bull. Volcanol. 24, 158-167.

Blatchford, A. (1953). Geology of the Tavua goldfield, Viti Levu, Fiji. Proc. Aust. Inst. Min. Met. Nos. 168-9, 13-51.

Blow, W.H. (1969). Late Middle Eocene to Recent planktonic foraminiferal biostratigraphy. Proc. First Internat. Conf. Planktonic Microfossils, Geneva, 1967, pp. 199-422. E.J. Brill.

Bonatti, E. and Arrhenius G. (1970). Acidic rocks on the Pacific Ocean floor. In: The Sea (ed. A.E. Maxwell), Vol. 4, Pt. I, pp. 454-464. Wiley Interscience.

Boyd, F.R. and England, J.L. (1959). Pyrope. Carnegie Inst. Year Bk. 58, 83-87.

Brock, R.W. (1924). Sketch of the geology of Viti Levu, Great Fiji. Proc. Roy. Soc. Canada 18. 63-83.

Brooks, C.K. (1970). Concentrations of zirconium and hafnium in some igneous and metamorphic rocks and minerals. Geochim. Cosmochim. Acta 34, 411-416.

Brothers, R.N. (1970). Petrochemical affinities of volcanic rocks from the Tonga-Kermadec island arc, Southwest Pacific. Bull. Volcanol. 34, 308-329.

Brothers, R.N. (1972). Tertiary orogenesis and high pressure metamorphism in New Caledonia. In: The Western Pacific: Island Arcs, Marginal Seas, and Geochemistry (ed. P. Coleman). W. Austral. Univ. Press.

———— and Searle, E.J. (1970). The geology of Raoul Island, Kermadec Group, Southwest Pacific. Bull. Volcanol. 34, 7-37.

Brown, D.A., Campbell, K.S.W. and Crook, K.A.W. (1968). The geological evolution of Australia and New Zealand. Pergamon Press, 409 pp.

Brown, G.C. and Fyfe, W.S. (1970). The production of granitic melts during ultrametamorphism. Contr. Mineral. Petrol. 28, 310-318.

Bryan, W.B. (1968). Low-potash dacite drift pumice from the Coral Sea. Geol. Mag. 105, 431-439.

Bultitude, R.J. and Green, D.H. (1971). Experimental study of crystal-liquid relationships at high pressures in olivine nephelinite and basanite compositions. J. Petrology 12, 121-147.

Burnham, C.W. (1967). Hydrothermal fluids at the magmatic stage. In: Geochemistry of Hydrothermal Ore Deposits (ed. H.L. Barnes), pp. 34-76. Holt, Rinehart and Winston, Inc.

Burns, R.G. (1970). Mineralogical applications of crystal field theory. Cambridge Univ. Press, 224 pp.

Byers, F.M.Jr. (1959). Geology of Umnak and Bogoslof Islands, Aleutian Islands. U.S. Geol. Surv. Bull. 1028-L.

———— (1961). Petrology of three volcanic suites, Umnak and Bogoslof Islands, Aleutian Islands, Alaska. Bull. Geol. Soc. Am. 72, 93-128.

Cambel, B. and Jarkovský, J. (1969). Geochemistry of pyrrhotite of various genetic types. Komenský Univ.: Bratislava, 333 pp.

Cann, J.R. (1969). Spilites from the Carlsberg Ridge, Indian Ocean. J. Petrology 10, 1-19.

———— (1970). Rb, Sr, Y, Zr and Nb in some ocean floor basaltic rocks. Earth Planet. Sci. Letters 10, 7-11.

———— (1971). Major element variations in ocean-floor basalts. Phil. Trans. Roy. Soc. Lond. A. 268, 495-505.

Cannon, H.L. and Hopps, H.C. (1971). Environmental geochemistry in health and disease. Geol. Soc. Am. Mem. 123.

Carey, S.W. (1963). The assymetry of the Earth. Aust. J. Sci. 25, 369-383 and 479-488.

———— (1970). New Guinea and Melanesia in the current revolution in concepts of the evolution of the Earth. Search 1, 178-189.

Carmichael, I.S.E. (1964). The petrology of Thingmuli, a Tertiary volcano in eastern Iceland. J. Petrology 5, 435-460.

———— (1967a). The iron-titanium oxides of volcanic rocks and their associated ferromagnesian silicates. Contr. Mineral. and Petrol. 14, 36-64.

Carmichael, I.S.E. (1967b). The mineralogy and petrology of the volcanic rocks from the Leucite Hills, Wyoming. Contr. Mineral. and Petrol. 15, 24-66.

———— and Nicholls, J. (1967). Iron-titanium oxides and oxygen fugacities in volcanic rocks. J. Geophys. Res. 72, 4665-4687.

Chamberlin, R.T. (1935). Review of Geology of Vitilevu, by H.S. Ladd. J. Geol. 43, 781-783.

Chase, C.G. (1971). Tectonic history of the Fiji Plateau. Bull. Geol. Soc. Am. 82, 3087-3110.

Chase, T.E, Karig, D.E., Newhouse, D.A. and Taylor, I.L. (1968). Bathymetry of the southwest Pacific. (Map). Scripps Inst. Oceanography.

Chayes, F. (1964a). Variance-covariance relations in some published Harker diagrams of volcanic suites. J. Petrology 5, 219-237.

———— (1964b). A petrographic distinction between Cenozoic volcanics in and around the open oceans. J. Geophys. Res. 69, 1573-88.

———— (1966). Alkaline and subalkaline basalts. Am. J. Sci. 264, 128-145.

Cherdyntsev, V.V., Malyshev, Z.A., Sokolova, I.N., Kazachevskii and Borisov, I.V. (1964). Isotopic composition of uranium and thorium in the supergene zone. Geochem. Internat. 3, 398-401.

———— and Senina (1970). Radioactivity of young volcanic rocks of Pacific islands. Geochemistry 7, 652-663.

Cherry, R.D., Hobbs, J.B.M., Erlank, A.J. and Willis, J.P. (1970). Thorium, Uranium, Potassium, Lead, Strontium, and Rubidium in silicate rocks by gamma-spectroscopy and/or X-ray fluorescence. Can. Spectroscopy 15, 1-8.

Christensen, N.I. (1970). Composition and evolution of the oceanic crust. Marine Geol. 8, 139-154.

Christiansen, R.L. and Lipman, P. (1972). Cenozoic volcanism and plate tectonic evolution of western United States, pt. 2, late Cenozoic. Phil. Trans. Roy. Soc. Lond. A 271, 249-284.

Christoffel, D.A. and Ross, D.I. (1970). A fracture zone in the southwest Pacific basin south of New Zealand and its implications for sea floor spreading. Earth Planet. Sci. Letters 8, 125-130.

Church, S.E. (1971). Strontium isotope and alkali element geochemistry of selected sediments from the northeast Pacific Ocean. Geochim. Cosmochim. Acta 35, 1300-06.

———— and Tilton, G.R. (in prep.). Lead and strontium isotope geochemistry of the Cascade mountains and their bearing on the genesis of andesitic magma.

Clark, S.P.Jr. and Ringwood, A.E. (1964). Density distribution and constitution of the mantle. Rev. Geophysics 2, 35-88.

Clarke, W.J. and Blow, W.H. (1969). The inter-relationships of some Late Eocene, Oligocene, and Miocene larger foraminifera and planktonic biostratigraphic indices. Proc. First Internat. Conf. Planktonic Microfossils, Geneva, 1967, pp. 82-97. E.J. Brill.

Cleary, J.R. and Simpson, D.W. (1971). Seismotectonics of the Australian continent. Nature 230, 239-241.

Coats, R.R. (1962). Magma type and Crustal Structure in the Aleutian arc. In: Crust of the Pacific Basin, Geophys. Mon. 6, 92-109.

Cohen, E.M. (1962). Revised geology of the Tavua goldfield. Aust. Inst. Min. Met., No. 204, 135-160.

Cole, J.W. (1967). Quaternary volcanic zones in the North Island of New Zealand and their relationship to structure. Bull. Volcanol. 30, 189-197.

Cole, W.S. (1960). Upper Eocene and Oligocene larger foraminifera from Viti Levu, Fiji. U.S. Geol. Surv. Prof. Pr. 374-A.

Coleman, P.J. (1966). The Solomon Islands as an island arc. Nature 211, 1249-51.

———— (1969). Derived Eocene larger foraminifera on Maewo, Eastern New Hebrides, and their south-west Pacific implications. 1967 Ann. Report New Hebrides Geol. Surv., pp. 36-37.

———— (1970). Geology of the Solomon and New Hebrides Islands, as part of the Melanesian Re-entrant, southwest Pacific. Pacific Science 24, 289-314.

Coleman, R.G. (1967). Glaucophane schists from California and New Caledonia. Tectonophysics 4, 479-498.

———— (1971). Plate tectonic emplacement of upper mantle peridotites along continental edges. J. Geophys. Res. 76, 1212-22.

———— , Lee, D.E., Beatty, L.B. and Brannock, W.W. (1965). Eclogites and eclogites: their differences and similarities. Bull. Geol. Soc. Am. 76, 483-508.

Colley, H. and Warden, A.J. (in prep.). The petrology of the New Hebrides.

Compston, W. Chappell, B.W., Arriens, P.A. and Vernon, M.J. (1970). The chemistry and age of Apollo 11 lunar material. Proc. Apollo 11 Lunar Sci. Conf. pp. 1007-28.

Coombs, D.S. (1963). Trends and affinities of basaltic magmas and pyroxenes as illustrated on the diopside-olivine-silica diagram. Mineralog. Soc. Am. Spec. Paper 1, 227-250.

———— and Wilkinson, J.F.G. (1969). Lineages and fractionation trends in unsaturated volcanic rocks from the East Otago Volcanic Province (New Zealand) and related rocks. J. Petrology 10, 440-501.

Cooper, J.A. (1963). The flame photometric determination of potassium in geological materials used for potassium-argon dating. Geochim. Cosmochim. Acta 27, 525-546.

Coulson, F.I.E. (1971). The geology of western Vanua Levu. Bull. Geol. Surv. Fiji, 17.

Crickmay, G.W., Ladd, H.S. and Hoffmeister, J.E. (1941). Shallow-water Globigerina sediments. Bull. Geol. Soc. Am. 52, 79-106.

Crook, K.A. (1963). Burial metamorphic rocks from Fiji. N.Z. J1. Geol. Geophys. 6, 681-704.

Cullen, D.J. (1970). A tectonic analysis of the South-west Pacific. N.Z. J1. Geol. Geophys. 13, 7-20.

Dalrymple, G.B. and Lanphere, M.A. (1969). Potassium-argon dating. W.H. Freeman and Co., 258 pp.

Davies, H.L. (1971). Peridotite-gabbro-basalt complex in eastern Papua: an overthrust plate of oceanic mantle and crust. Austral. Bur. Min. Res. Bull. 128.

——— and Smith, I.E. (1971). Geology of eastern Papua. Bull. Geol. Soc. Am. 82, 3299-3312.

Davis, W.M. (1927). A migrating anticline in Fiji. Am. J. Sci. 14, 333-351.

Denham, D. (1969). Distribution of earthquakes in the New Guinea-Solomon Islands region. J. Geophys. Res. 74, 4290-99.

Denholm, L.S. (1966). Structural and economic aspect of the Vatukoula caldera, Fiji. Bull. Volcanol. 29, 223-233.

——— (1967). Lode structures and ore shoots at Vatukoula, Fiji. Aust. Inst. Min. Met., No. 222, 73-83.

Derrick, R.A. (1965). The Fiji Islands, a geographical handbook. Suva: Government Press. 336 pp.

Dewey, J.F. and Bird, J.M. (1970). Mountain belts and the New Global Tectonics. J. Geophys. Res. 75, 2625-47.

Dickinson, W.R. (1966). Petrography of specimens from the Mamanutha Group. Geol. Surv. Fiji Note 20/66.

——— (1967). Tectonic development of Fiji. Tectonophysics 4, 543-553.

——— (1968a). Circum-Pacific andesite types. J. Geophys. Res. 73, 2261-69.

——— (1968b). Sedimentation of volcaniclastic strata of the Pliocene Koroimavua Group in northwest Viti Levu, Fiji. Am. J. Sci. 266, 440-453.

——— (1970a). Relations of andesites, granites, and derivative sandstones to arc-trench tectonics. Rev. Geophys. Space Phys. 8, 813-860.

——— (1970b). Relations of andesitic volcanic chains and granitic batholith belts to the deep structures of orogenic arcs. Geol. Soc. Lond. No. 1662.

——— (1971a). Detrital modes of New Zealand greywackes. Sediment. Geol. 5, 37-56.

——— (1971b). Plate tectonic models for orogeny at continental margins. Nature 232, 41-42.

——— (1971c). Plate tectonics in geologic history. Science 174, 107-113.

——— (1972). Reconstruction of past arc-trench systems from petrotectonic assemblages in the island arcs of the western Pacific. In: The Western Pacific: Island Arcs, Marginal Seas, and Geochemistry (ed. P. Coleman). W. Austral. Univ. Press.

——— and Luth, W.C. (1971). A model for plate tectonic evolution of mantle layers. Science 174, 400-404.

Dickinson, W.R., Rickard, M.J., Coulson, F.I., Smith, J.G. and Lawrence R.L. (1968). Late Cenozoic Shoshonitic lavas in northwestern Viti Levu, Fiji. Nature 219, 148.

Duncan, A.R. and Taylor, S.R. (1969). Trace element analyses of magnetites from andesitic and dacitic lavas from Bay of Plenty, New Zealand. Contr. Mineral. and Petrol. 20, 30-33.

Eames, F.E., Banner, F.T. and Blow, W.H. (1958). London paleontological note No. 171 (British Petroleum Co. Ltd.) on fifteen samples from the Fiji islands. Geol. Surv. Fiji Rept. 42.

Ehrlich, E.N. (1968). Petrochimiya Kainozoiskoi Kurilo-Kamchatskoi vulkanitsheskoi provincii. Nauka Moskva.

Elsasser, W.M. (1971). Sea-floor spreading as thermal convection. J. Geophys. Res. 76, 1101-12.

Engel, A.E.J., Engel, C.G. and Havens, R.G. (1965). Chemical characteristics of oceanic basalts and the upper mantle. Bull. Geol. Soc. Am. 76, 719-734.

Ernst, W.G. (1971). Metamorphic zonations on presumably subducted lithospheric plates from Japan, California, and the Alps. Contr. Mineral. and Petrol. 34, 43-59.

Essene, E.J., Hensen, B.J., and Green, D.H. (1970). Experimental study of amphibolite and eclogite stability. Phys. Earth Planet. Interiors 3, 378-384.

Evernden, J.F. and Evernden, R.K.S. (1970). The Cenozoic time scale. Geol. Soc. Am. Spec. Pr. 124, 71-90.

*Ewart, A., Bryan, W.B. and Graham, A.L. (1972). The petrology and geochemistry of the igneous rocks from Eua, Tonga. Bull. Geol. Soc. Am. 83.

———— and Stipp, J.J. (1968). Petrogenesis of the volcanic rocks of the Central North Island, New Zealand, as indicated by a study of Sr 87/86 ratios, and Sr, Rb, K, U, and Th abundances. Geochim. Cosmochim. Acta 32, 699-736.

Fairbairn, H.W. and Hurley, P.M. (1971). Evaluation of X-ray fluorescence and mass spectrometric analyses of Rb and Sr in some silicate standards. Geochim. Cosmochim. Acta 35, 149-156.

Fairbridge, R.W. (1961). The Melanesian Border Plateau, a zone of crustal shearing in the S.W. Pacific. Publ. Bur. Cent. Seism. Internat., I.U.G.G., series A, No. 22, pp. 137-149.

———— and Stewart, H.B.Jr. (1960). Alexa Bank, a drowned atoll on the Melanesian Border Plateau. Deep-Sea Res. 7, 100-116.

Farberov, A.I. and Gorelchik, V.I. (1971). Anomalous seismic effect under volcanoes and some features of deep-seated structure of volcanic areas. Bull. Volcanol. 35, 212-224.

Fedotov, S.A. (1968). On deep structure, properties of the upper mantle, and volcanism of the Kuril-Kamchatka island arc according to seismic data. In: The Crust and Upper Mantle of the Pacific Area (eds. L. Knopoff, C.L. Drake, and P.J. Hart), A.G.U. Mon. 12, 131-139.

Fein, C.D. (1971). Geochemical variations across an island arc. Abs. Geol. Soc. Am. Ann. Mtgs. 3, 563.

*Ewart, A. and Bryan, W.B. (1972). Petrology and geochemistry of the Tongan Islands. In: The Western Pacific: Island Arcs, Marginal Seas, and Geochemistry (ed. P. Coleman). W. Austral. Univ. Press.

Fitch, T.J. and Scholz C.H. (1971). Mechanism of underthrusting in southwest Japan: a model of convergent plate interactions. J. Geophys. Res. 76, 7260-92.

Fitton, J.G. (1971). The generation of magmas in island arcs. Earth Planet. Sci. Letters 11, 63-67.

Fleischer, M. (1955). Minor elements in some sulfide minerals. Econ. Geol. 50th Ann. Vol., pp. 970-1024.

Fleming, C.A. (1970). The Mesozoic of New Zealand: chapters in the history of the circum-Pacific mobile belt. Q. Jl. geol. Soc. Lond. 125, 125-170.

Flower, M.F.J. (1971). Evidence for the role of phlogopite in the genesis of alkali basalts. Contr. Mineral. and Petrol. 32, 126-137.

Forbes, R.B., Ray, D.K., Katsura, T., Matsumoto, H., Haramura, H. and Furst, M.J. (1969). The comparative chemical composition of continental vs. island arc andesites in Alaska. In: Proc. Andesite Conf. (ed. A.R. McBirney), Dept. Geol. Min. Res. Oregon Bull. 65, pp. 111-120.

Foye, W.G. (1918). Geological observations in Fiji. Proc. Am. Acad. Arts Sci. 54, 1-145.

Frey, F.A., Haskin, M.A., Pbetz, J.A. and Haskin, L.A. (1968). Rare earth abundances in some basic rocks. J. Geophys. Res. 73, 6085-98.

Funnell, B.M. (1964). The Tertiary period. In: The Phanerozoic Time-scale (eds. Harland, W.B., Smith, A.G., and Wilcock, B.), Geol. Soc. Lond. Quart. Jour., 120S, 179-91.

Gass, I.G., Harris, P.G. and Holdgate, M.W. (1963). Pumice eruption in the area of the South Sandwich Islands. Geol. Mag. 100, 321-330.

Gast, P.W. (1968). Trace element fractionation and the origin of tholeiitic and alkaline magma types. Geochim. Cosmochim. Acta 32, 1057-86.

Gill, J.B. (1970). Geochemistry of Viti Levu, Fiji, and its evolution as an island arc. Contr. Mineral. Petrol. 27, 179-203.

———— and Murthy, V.R. (1970). Distribution of K, Rb, Sr, and Ba in Nain anorthosite plagioclase. Geochim. Cosmochim. Acta 34, 401-408.

Glaessner, M.F. (1950). Geotectonic position of New Guinea. Bull. Am. Ass. Petr. Geol. 34, 856-881.

Gorshkov, G.S. (1962). Petrochemical features of volcanism in relation to the types of the earth's crust. In: Crust of the Pacific Basin, A.G.U. Mon. 6, 110-115.

———— (1970). Volcanism and the Upper Mantle. Plenum Press. 385 pp.

Graham, A.L. and Nicholls, G.D. (1969). Mass spectrographic determinations of lanthanide element contents in basalts. Geochim. Cosmochim. Acta 33, 555-568.

Gray, C.M. (1971). Strontium isotopic studies on granulites. Unpub. Ph.D. thesis, A.N.U.

Green, D.H. (1970). The origin of basaltic and nephelinitic magmas. Trans. Leicester Lit. Phil. Soc. 64, 28-54.

———— (1971). Composition of basaltic magmas as indicators of conditions of origin: application to oceanic volcanism. Phil. Trans. Roy. Soc. Lond. A 268, 707-725.

———— (1972). Magmatic activity as the major process in chemical evolution of the Earth's crust and mantle. Tectonophysics (in press).

———— and Ringwood, A.E. (1967). The genesis of basaltic magmas. Contr. Mineral. and Petrol. 15, 103-190.

———— and Ringwood, A.E. (1970). Mineralogy of peridotitic compositions under upper mantle conditions. Phys. Earth Planet. Interiors 3, 359-371.

———— and Ringwood, A.E. (1972). A comparison of recent experimental data on the gabbro-garnet granulite-eclogite transition. (preprint)

Green, T.H. (1967). High pressure experimental investigations on the origin of high-alumina basalt, andesite, and anorthosite. Unpub. Ph.D. thesis, A.N.U.

———— (1972). Crystallization of calc-alkaline andesite under controlled high-pressure hydrous conditions. Contr. Mineral. and Petrol. 34 (in press).

———— and Ringwood, A.E. (1968). Genesis of the calc-alkaline igneous rock suite. Contr. Mineral. and Petrol. 18, 105-162.

Griffin, W.L. and Murthy, V.R. (1969). Distribution of K, Rb, Sr and Ba in some minerals relevant to basalt genesis. Geochim. Cosmochim. Acta 33, 1389-1414.

Griffiths, J.R. (1971a). Continental margin tectonics and the evolution of South East Australia. The APEA Journal, 1971, 75-79.

———— (1971b). Reconstruction of the South-west Pacific margin of Gondwanaland. Nature 234, 203-207.

Groube, L.M. (1971). Tonga, lapita pottery, and Polynesian origins. J. Polynesian Soc. 80, 278-316.

Gunn, B.M. and Mooser, F. (1970). Geochemistry of the volcanics of central Mexico. Bull. Volcanol. 24, 577-616.

Hackman, B.D. (1972). The Solomons fractured arc. In: The Western Pacific: Island Arcs, Marginal Seas, and Geochemistry (ed. P. Coleman). W. Austral. Univ. Press.

Hakli, T. and Wright, T.L. (1967). The fractionation of nickel between olivine and augite as a geothermometer. Geochim. Cosmochim. Acta 31, 877-884.

Hamilton, D.L. and Anderson, G.M. (1968). Effects of water and oxygen pressures on the crystallization of basaltic magmas. In: Basalts (eds. H.H. Hess and A. Poldevaart), 1, 445-482. Interscience Pub.

Hamilton, W. (1969). The volcanic central Andes - a modern model for the Cretaceous batholiths and tectonics of western North America. In: Proc. Andesite Conference (ed. A.R. McBirney). Dept. Geol. Min. Res. Oregon Bull. 65, 175-184.

Hamilton, W. (1970). Tectonic map of Indonesia - a progress report. U.S. Geol. Surv. Report (IR) IND-5.

———— and Myers, W.B. (1967). The nature of batholiths. U.S. Geol. Surv. Prof. Pr. 554-C, pp. 1-26.

Hammaguchi, H. and Kuroda, R. (1969). Tin. In: Handbook of Geochemistry (ed. K.H. Wedepohl). Springer-Verlag. pp. 50-E-1-5.

Hanks, T.C. and Whitcomb, J.H. (1971). Comments on Paper by J.W. Minear and M.N. Toksöz, "Thermal Regime of a Down-going Slab and New Global Tectonics". J. Geophys. Res. 76, 613-616.

Hart, S.R. (1969). K, Rb, Cs contents and K/Rb, K/Cs ratios of fresh and altered submarine basalts. Earth Planet. Sci. Letters 6, 295-303.

———— (1971a). K, Rb, Cs, Sr, and Ba contents and Sr isotope ratios of ocean floor basalts. Phil. Trans. Roy. Soc. Lond. A 268, 573-587.

———— (1971b). The geochemistry of basaltic rocks. Carnegie Inst. Year Bk. 70, 353-355.

———— , Brooks, C., Krogh, T.E., Davis, G.L. and Nava, D. (1970). Ancient and modern volcanic rocks: a trace element model. Earth Planet. Sci. Letters 10, 17-28.

———— , Gunn, B.M. and Watkins, N.D. (1971). Intralava variation of alkali elements in Icelandic basalt. Am. J. Sci. 270, 315-318.

———— and Nalwalk, A.J. (1970). K, Rb, Cs and Sr relationships in submarine basalts from the Puerto Rico trench. Geochim. Cosmochim. Acta 34, 145-156.

Hasebe, K., Fujii, N. and Uyeda, S. (1970). Thermal processes under island arcs. Tectonophysics 10, 335-355.

Haskin, L.A., Allen, R.O., Helmke, P.A., Paster T.P., Anderson, M.R., Korotev, R.L. and Zweifel, K.A. (1970). Rare earths and other trace elements in Apollo 11 lunar samples. Proc. Apollo 11 Lunar Sci. Conf., pp. 1213-32.

Hatherton, T.H. (1971). Plate consumption: some comments from an island arc. Nature 234, 294-5.

———— and Dickinson, W.R. (1969). The relationship between andesitic volcanism and seismicity in Indonesia, the Lesser Antilles, and other island arcs. J. Geophys. Res. 74, 5301-10.

Hawkins, J.W. and Nishimori, R. (1971). Ocean ridge-type tholeiitic rocks from the Lau Basin: their petrology and significance. Abs. Geol. Soc. Am. Ann. Mtg. 3, 594.

Hayes, D.E. and Ewing, M. (1970). Pacific boundary structure. In: The Sea (ed. A.E. Maxwell), Vol. 4, part II, pp. 29-72. Wiley Interscience.

———— and Ringis, J. (1972). The early opening of the Central Tasman Sea. Abs. Int'l. Sym. Oceanography of the South Pacific. p. 65.

Hedge, C.E. (1971). Nickel in high-alumina basalts. Geochim. Cosmochim. Acta 35, 522-524.

Hedge, C.E. and Knight, R.J. (1969). Lead and strontium isotopes in volcanic rocks from Northern Honshu, Japan. Geochem. J. 3, 15-24.

———— and Lewis, J.F. (1971). Isotopic composition of strontium in three basalt-andesite centers along the Lesser Antilles Arc. Contr. Mineral. and Petrol. 32, 39-47.

———— and Peterman, Z.E. (1970). The strontium isotopic composition of basalts from the Gordo and Juan de Fuca Rises, Northeastern Pacific Ocean. Contr. Mineral. and Petrol. 27, 114-120.

Heier, K.S. (1965). Radioactive elements in the continental crust. Nature 208, 479-480.

———— , McDougall, I. and Adams, J.A.S. (1964). Thorium, uranium, and potassium concentrations in Hawaiian lavas. Nature 201, 254-256.

———— and Rogers, J.J.W. (1963). Radiometric determination of thorium, uranium, and potassium in basalts and in two magmatic differentiation series. Geochim. Cosmochim. Acta 27, 137-154.

———— and Thoresen, K. (1971). Geochemistry of high grade metamorphic rocks, Lofoten-Vesterålen, North Norway. Geochim. Cosmochim. Acta 35, 89-100.

Hekinian, R. (1971). Chemical and mineralogical differences between abyssal hill basalts and ridge tholeiites in the eastern Pacific Ocean. Marine Geol. 11, 77-91.

Herrmann, A.G. and Wedepohl, K.H. (1970). Untersuchungen an spilitischen Gesteinen der variskischen Geosynkline in Nordwestdeutschland. Contr. Mineral. and Petrol. 29, 255-274.

Hess, H.H. and Maxwell, J.C. (1949). Major structural features of the southwest Pacific: a preliminary interpretation of H.O.5484, bathymetric chart, New Guinea to New Zealand. Proc. Pacific Sci. Congr. Pacific Sci. Assoc., 7th, New Zealand, 1949, 2, 14-17.

Higuchi, H. and Nagasawa, H. (1969). Partition of trace elements between rock-forming minerals and the host volcanic rocks. Earth Planet. Sci. Letters 7, 281-287.

Hill, R.E.T. and Boettcher, A.L. (1970). Water in the Earth's mantle: melting curves of basalt-water and basalt-water-carbon dioxide. Science 167, 980-981.

Hindle, W.H. (1970). Geochemistry of samples from Vanua Levu and other islands in the Fiji Group. Unpub. M.Sc. thesis, Leeds Univ.

———— (in prep.). Geology of mid-western Vanua Levu. Bull. Geol. Surv. Fiji, 19.

Hirst, J.A. (1965). Geology of East and Northeast Viti Levu. Bull. Geol. Surv. Fiji, 12.

Hoffmeister, J.E. (1932). Geology of Eua, Tonga. B.P. Bishop Mus. Bull. 96.

Holloway, J.R. and Burnham, C.W. (1972). Melting relations of basalt with equilibrium water pressures less than total pressure. J. Petrology.

Horibe, Y. (ed.) (1970). Preliminary report of the Hakuho Maru Cruise KH-68-4. Tokyo: Ocean Res. Inst., 170 pp.

Houtz, R.E. (1960). Geology of Singatoka area, Viti Levu. Bull. Geol. Surv. Fiji, 6.

——— (1963). Regional geology - Keiyasi area. Bull. Geol. Surv. Fiji, 10.

Hubbard, N.J. (1969). A chemical comparison of oceanic ridge, Hawaiian tholeiitic, and Hawaiian alkalic basalts. Earth Planet. Sci. Letters 5, 346-352.

——— (1971). Some chemical features of lavas from the Manu'a Islands, Samoa. Pacific Sci. 25, 178-187.

Hunkins, K. and Kuo, J.T. (1965). Surface wave dispersion in the Tonga-Fiji region. Bull. Seis. Soc. Am. 55, 135-145.

Hurley, P.M. and Rand, J.R. (1969). Pre-drift continental nuclei. Science 164, 1229-42.

Ibbotson, P. (1960a). Geology of the Suva area, Viti Levu. Bull. Geol. Surv. Fiji, 4.

——— (1960b). Geological reconnaissance of Taveuni. Geol. Surv. Fiji. Long Report 63.

——— (1961). The geology of Ovalau, Moturiki, and Naingani. Bull. Geol. Surv. Fiji, 9.

——— (1962). The geology of the Tavua area, Viti Levu. Bull. Geol. Surv. Fiji, 8.

——— (1967). Petrology of the Tertiary Caldera, Tavua goldfield. Mem. Geol. Surv. Fiji, 3.

——— (1969). The geology of east-central Vanua Levu. Bull. Geol. Surv. Fiji, 16.

Iddings (1899). Geology of the Yellowstone National Park, part II. U.S. Geol. Surv. Mon. 32, chapter 9.

Irvine, T.N. and Baragar, W.R.A. (1971). A guide to the chemical classification of the common volcanic rocks. Canad. J. Earth Sci. 8, 523-548.

Irving, A. (1972). Geochemical and high-pressure experimental studies of xenoliths, megacrysts and basalts from southeastern Australia. Unpub. Ph.D. thesis, A.N.U.

Isacks, B. and Molnar, P. (1971). Distribution of stresses in the descending lithosphere from a global survey of focal-mechanism solutions of mantle earthquakes. Rev. Geophys. and Space Phys. 9, 103-174.

——— , Oliver, J. and Sykes, L.R. (1968). Seismology and the New Global Tectonics. J. Geophys. Res. 73, 5855-99.

——— , Sykes, L.R. and Oliver, J. (1969). Focal mechanisms of deep and shallow earthquakes in the Tonga-Kermadec region and the tectonics of island arcs. Bull. Geol. Soc. Am. 80, 1443-70.

Jakes, P. (1970). Analytical and experimental geochemistry of volcanic rocks from island arcs. Unpub. Ph.D. thesis, A.N.U.

——— and Gill, J. (1970). Rare earth elements and the island arc tholeiitic series. Earth Planet. Sci. Letters 9, 17-28.

Jakes, P. and Smith, I.E. (1970). High potassium calc-alkaline rocks from Cape Nelson, Eastern Papua. Contr. Mineral and Petrol. 28, 259-271.

———— and White, A.J.R. (1969). Structure of the Melanesian Arcs and correlation with distribution of magma types. Tectonophysics 8, 223-236.

———— and White, A.J.R. (1970). K/Rb ratios of rocks from island arcs. Geochim. Cosmochim. Acta 34, 849-856.

———— ———— (1971). Composition of island arcs and continental growth. Earth Planet. Sci. Letters 12, 224-230.

———— ———— (1972). Major and trace element abundances in volcanic rocks of orogenic areas. Bull. Geol. Soc. Am. 83, 29-40.

Johannsen, A. (1937). A Descriptive Petrography of the Igneous Rocks. U. Chicago Press, 4 vols.

Johnson, J.H. and Ferris, B.J. (1950). Tertiary and Pleistocene coralline algae from Lau, Fiji. B.P. Bishop Mus. Bull. 201.

Johnson, R.W. (1970). Volcanoes of New Britain: a study of preliminary petrological data. Austral. Bur. Min. Res. Record 1970/72.

———— , Mackenzie, D.E. and Smith, I.E. (1971). Seismicity and Late Cenozoic volcanism in parts of Papua-New Guinea. Tectonophysics 12, 15-22.

Johnson, T. and Molnar, P. (1972). Focal mechanisms and plate tectonics of the southwest Pacific. (preprint)

Jolly, W. (1971). Potassium-rich igneous rocks from Puerto Rico. Bull. Geol. Soc. Am. 82, 399-408.

Joplin, G.A. (1968). The Shoshonite association: a review. J. Geol. Soc. Aust. 15, 275-294.

Karig, D.E. (1970). Ridges and basins of the Tonga-Kermadec Island Arc system. J. Geophys. Res. 75, 239-254.

———— (1971a). Structural history of the Mariana island arc system. Bull. Geol. Soc. Am. 82, 323-344.

———— (1971b). Origin and development of marginal basins in the western Pacific. J. Geophys. Res. 76, 2542-61.

Kay, R., Hubbard, N.J. and Gast, P.W. (1970). Chemical characteristics and origin of oceanic ridge volcanic rocks. J. Geophys. Res. 75, 1585-1614.

Kesson, S. and Price, R.C. (1972). The major and trace element chemistry of kaersutite and its bearing on the petrogenesis of alkaline rocks. Nature (in press).

Kiss, E. (1967). Chemical determination of some major constituents in rocks and minerals. Anal. Chim. Acta 39, 223-234.

Kleeman, A.W. (1967). Sampling error in the chemical analysis of rocks. J. Geol. Soc. Aust. 14, 43-47.

Kleeman, J.D. (1970). Fisson track studies and the geochemistry of uranium in rocks from the crust and upper mantle. Unpub. Ph.D. thesis, A.N.U.

Kleinpell, R.M. (1954). Neogene smaller foraminifera from Lau, Fiji. B.P. Bishop Museum Bull. 211.

Korringa, M.K. and Noble, D.C. (1970). Feldspar-liquid partition coefficients for Sr and Ba based on a Rayleigh fractionation model. Abs. Geol. Soc. Am. Ann. Mtg. 2, 598.

Kravchenko, S.M. and Semenov, G.G. (1970). Gamma-spectrometric determination of uranium, thorium, and potassium in Cenozoic basalt, andesite, and dacite from the Kurile Islands and the Lesser Caucasus. Dokl. Akad. Nauk. USSR. 193, 437-40.

Kroenke, L.W. (1970). The bathymetry and surficial geology of the Ontong Java Plateau. H.I.G. Ann. Prog. Rep. 1970, 64.

Kuno, H. (1950). Petrology of Hakone volcano and the adjacent areas, Japan. Bull. Geol. Soc. Am. 61, 957-1020.

———— (1959). Origin of Cenozoic petrographic provinces of Japan and surrounding areas. Bull. Volcanol. 20, 37-76.

———— (1960). High-alumina basalt. J. Petrology 1, 121-145.

———— (1962). Catalogue of the active volcanoes of the world. Part 11, Japan, Taiwan, and Marianas. Int. Assoc. Volcanology.

———— (1966). Lateral variation of basalt magma across continental margins and island arcs. Bull. Volcanol. 29, 195-222.

———— (1968a). Origin of andesite and its bearing on the island arc structure. Bull. Volcanol. 32, 141-176.

———— (1968b). Differentiation of basalt magmas. In: Basalts (eds. H.H. Hess and A. Poldevaart), 2, 624-688. Interscience Pub.

Kurasawa, H. (1970). Strontium and lead isotopes of volcanic rocks in Japan. In: Recent Developments in Mass Spectrometry (eds. K. Ogata and T. Hayakawa). U. Tokyo Press, 666-670.

Kurentsova, N.A. and Shreyder, A.A. (1971). Geomorphological and geomagnetic investigations in the North Fiji Basin. Oceanology 11, 54-57.

Kushiro, I. (1970). Systems bearing on melting of the upper mantle under hydrous conditions. Carnegie Inst. Washington Year Bk. 68, 240-244.

———— , Yoder, H.S. and Nishikawa, M. (1968). Effect of water on the melting of enstatite. Bull. Geol. Soc. Am. 79, 1685-92.

Lachenbruch, A.H. (1968). Preliminary geothermal model of the Sierra Nevada. J. Geophys. Res. 73, 6977-90.

———— (1971). Vertical gradients of heat production in the continental crust 1: theoretical detectability from near-surface measurements. J. Geophys. Res. 76, 3842-51.

Ladd, H.S. and Hoffmeister, J.E. (1945). Geology of Lau, Fiji. B.P. Bishop Museum Bull. 181.

Lambert, I.B. and Heier, K.S. (1967). The vertical distribution of uranium, thorium, and potassium in the continental crust. Geochim. Cosmochim. Acta 31, 377-390.

Lambert, I.B. and Heier, K.S. (1968). Geochemical investigations of deep-seated rocks in the Australian shield. Lithos 1, 30-53.

———— and Wyllie, P.J. (1970). Low-velocity zone of the Earth's mantle: incipient melting caused by water. Science 169, 764-6.

Leonova, L.L. and Udaltzova, N.I. (1970). U, Th, Li, Rb, Cs in the volcanic rocks of the Kurile Isles and Kamchatka. Geokhimya no. 11, 1329-34.

Le Pichon, X. (1968). Sea-floor spreading and continental drift. J. Geophys. Res. 73, 3661-97.

———— and Heirtzler, J.R. (1968). Magnetic anomalies in the Indian Ocean and sea-floor spreading. J. Geophys. Res. 73, 2101-17.

Levi, B. (1969). Burial metamorphism of a Cretaceous volcanic sequence west from Santiago, Chile. Contr. Mineral. and Petrol. 24, 30-49.

Lillie, A.R. and Brothers, R.N. (1970). The geology of New Caledonia. N.Z. J1. Geol. Geophys. 13, 145-183.

Lindren, W. and Ransome, F.L. (1906). Geology and gold deposits of the Cripple Creek District, Colarado. U.S.G.S. Prof. Pr. 54.

Liou, J.G. and Ito, K. (1971). Experimental studies of the phase relations between greenschist and amphibolite in a basaltic system. Abs. Geol. Soc. Am. Ann. Mtg. 3, 634-635.

Lipman, P.W. (1970). Relations between Cenozoic andesitic and rhyolitic volcanism, western interior United States. Proc. Geol. Soc. Lond., No. 1662, 36-39.

———— , Prostka, H.J., and Christiansen, R.L. (1971). Evolving subduction zones in the western United States, as interpreted from igneous rocks. Science 174, 821-825.

Lowder, G.G. and Carmichael, I.S.E. (1969). The volcanoes and caldera of Talasea, New Britain. Bull. Geol. Soc. Am. 81, 17-38.

Macdonald, G.A. (1968). Contribution to the petrology of Tutuila, American Samoa. Geol. Rundsch. 57, 821-837.

McDougall, I. (1963). K-Ar ages of some rocks from Viti Levu, Fiji. Nature 198, 677.

McKenzie, D.P. (1969). Speculations on the consequences and causes of plate motions. Geophys. J. R. astr. Soc. 18, 1-32.

———— (1970). Temperature and potential temperature beneath island arcs. Tectonophysics 10, 357-366.

Maclean, W.H. (1969). Liquidus phase relationships in the FeS-FeO-Fe_3O_4-SiO_2 system, and their application in geology. Econ. Geol. 64, 865-884.

Macnab, R.P. (1970). Geology of the Gazelle Peninsula T.P.N.G. Austral. Bur. Min. Res. Record 1970/63.

Malahoff, A. (1970). Magnetic and bathymetric relations for two local areas in the Fiji Plateau - New Hebrides region. H.I.G. Rep. 70-27, pp. 65-66.

Malahoff, A. (in press). Gravity and magnetic studies of the New Hebrides island arc. Bull. Geol. Soc. Am.

———— (preprint). The Fiji Plateau: on active spiralling tectonic system.

———— and Woollard, G.P. (1969). The New Hebrides Islands gravity network. H.I.G. Rep. 69-14.

Mallick, D.I.J. (1972). The New Hebrides Archipelago, In: The Western Pacific: Island Arcs, Marginal Seas, and Geochemistry (ed. P. Coleman). W. Austral. Univ. Press.

Markham, N.L. (1960). Synthetic and natural phases in the system Au-Ag-Te, part II. Econ. Geol. 55, 1460-77.

Markhinan, E.K. (1968). Volcanism as an agent of formation of the Earth's crust. In: The Crust and Upper Mantle of the Pacific Area, A.G.U. Mon. 12, 413-422.

Marshall, P. (1912). Address by President. Rept. Australas. Assoc. Adv. Sci. 13, 90-99.

Mason, B. (1968). Kaersutite from San Carlos, Arizona, with comments on the paragenesis of this mineral. Mineral. Mag. 36, 997-1002.

Masuda, A. (1966). Lanthanides in basalts of Japan with three distinct types. Geochem. J. 1, 11-26.

———— (1967). Lanthanide concentration ratios between pyroxene and garnet. Earth Planet. Sci. Letters 3, 25-28.

———— (1968). Geochemistry of lanthanides in basalts of central Japan. Earth Planet. Sci. Letters 4, 284-292.

———— and Kushiro, I. (1970). Experimental determination of partition coefficients of ten rare earth elements and barium between clinopyroxene and liquid in the synthetic silicate system at 20 kilobar pressure. Contr. Mineral. and Petrol. 26, 42-49.

Matsumoto, T. (1971). Seismic body waves observed in the vicinity of Mount Katmai, Alaska, and evidence for the existence of molten chambers. Bull. Geol. Soc. Am. 82, 2905-20.

Matthews, D.H. (1971). Altered basalts from Swallow Bank, an abyssal hill in the N.E. Atlantic and from a nearby seamount. Phil. Trans. Roy. Soc. Lond. A 268, 551-572.

Maynard, G.L. (1970). Crustal layer of seismic velocity 6.9 to 7.6 kilometers per second under the deep oceans. Science 168, 120-121.

Melson, W.G., Jarosewich, E. and Lundquist, C.A. (1970). Volcanic eruption and Metis Shoal, Tonga, 1967-1968: description and petrology. Smithsonian Contr. Earth Sci. No. 4.

———— , Thompson, G. and van Andel, T. (1968). Volcanism and metamorphism in the Mid-Atlantic Ridge, 22°N Latitude. J. Geophys. Res. 73, 5925-43.

Meyer, H.O.A. and Brookins, D.G. (1971). Eclogite xenoliths from Stockdale Kimberlite, Kansas. Contr. Mineral. and Petrol. 34, 60-72.

Milson, J.S. (1970). The evolution of the Lau Ridge, Fiji Islands. Earth Planet. Sci. Letters 8, 258-60.

Minato, M., Gorai, M. and Hunahashi, M. (1965). The geologic development of the Japanese Islands. Tokyo: Tsukiji Shokan Co., 442 pp.

Minear, J.W. and Toksöz, M.N. (1970). Thermal regime of a downgoing slab and new global tectonics. J. Geophys. Res. 75, 1397-1419.

Mitchell, A.H.G. and Reading, H.G. (1969). Continental margins, geosynclines, and ocean floor spreading. J. Geol. 77, 629-646.

———— and Reading, H.G. (1971). Evolution of island arcs. J. Geol. 79, 253-284.

———— and Warden, A.J. (1971). Geological evolution of the New Hebrides island arc. Jour. Geol. Soc. 127, 501-529.

Mitronovas, W. and Isacks, B.L. (1971). Seismic velocity anomalies in the upper mantle beneath the Tonga-Kermadec island arc. J. Geophys. Res. 76, 7154-80.

Miyashiro, A., Shido, F. and Ewing, M. (1971). Metamorphism in the Mid-Atlantic Ridge near 24° and 30°N. Phil. Trans. Roy. Soc. Lond. A. 268, 589-604.

Moberly, R.Jr. (1972). Youthful oceanic lithosphere of marginal seas, western Pacific. In: The Western Pacific: Island Arcs, Marginal Seas, and Geochemistry (ed. P. Coleman). W. Austral. Univ. Press.

Moore, J.G. (1970). Water content of basalt erupted on the ocean floor. Contr. Mineral. and Petrol. 28, 272-279.

Morgan, W.J. (1968). Rises, trenches, great faults, and crustal blocks. J. Geophys. Res. 73, 1959-82.

Morgan, W.R. (1966). A note on the petrology of some lava types from East New Guinea. J. Geol. Soc. Aust. 13, 583-591.

Nagasawa, H. (1970). Rare earth concentration in zircon and apatite and their host dacites and granites. Earth Planet. Sci. Letters 5, 47-51.

———— and Schnetzler, C.C. (1971). Partioning of rare earth, alkali, and alkaline earth elements between phenocrysts and acidic igneous magma. Geochim. Cosmochim. Acta 35, 953-968.

Naldrett, A.J. (1969). A portion of the system Fe-S-O between 900 and 1080°C and its application to sulfide ore magmas. J. Petrology 10, 171-201.

Neuman van Padang, M. (1951). Catalogue of the active volcanoes of the world. Part 1, Indonesia. Int. Assoc. Volcanology.

Neumann, H., Mead, J. and Vitaliano, C.J. (1954). Trace element variation during fractional crystallization as calculated from the distribution law. Geochim. Cosmochim. Acta 6, 90-99.

Nicholls, I.A. (1971). Santorini Volcano, Greece - tectonic and petrochemical relationships with volcanics of the Aegean region. Tectonophysics 11, 377-385.

Nicholls, J. and Carmichael, I.S.E. (1969). A commentary on the Absarokite-Shoshonite-Banakite Series of Wyoming, U.S.A. Schweiz. Mineral. und Petrol. Mitt. 49, 47-64.

Norrish, K. and Chappell, B.W. (1967). X-ray fluorescence spectrography. In: Physical Methods in Determinative Mineralogy (ed. J. Zussman), pp. 161-214. Academic Press.

Norrish, K. and Hutton, J.T. (1969). An accurate X-ray spectrographic method for the analysis of a wide range of geological samples. Geochim. Cosmochim. Acta 33, 431-453.

Officer, Charles B. (1955). Southwest Pacific crustal structure. Trans. Am. Geophys. Un. 36, 449-459.

O'Hara, M.J. (1968). The bearing of phase equilibria studies in synthetic and natural systems on the origin and evolution of basic and ultrabasic rocks. Earth-Sci. Rev. 4, 69-134.

Oliver, J. and Isacks, B. (1967). Deep earthquake zones, anomalous structures in the upper mantle, and the lithosphere. J. Geophys. Res. 72, 4259-75.

Ono, K. (1962). Chemical composition of volcanic rocks in Japan. Geol. Surv. Japan, 441 pp.

Osborn, E.F. (1969). The complementariness of orogenic andesite and alpine peridotite. Geochim. Cosmochim. Acta 33, 307-324.

Oversby, B. (1971). Palaeozoic plate tectonics in the southern Tasman Geosyncline. Nature 234, 45-47.

Oversby, V.M. and Gast, P.W. (1970). Isotopic composition of lead from oceanic islands. J. Geophys. Res. 75, 2097-2114.

Oxburgh, E.R. and Turcotte, D.L. (1970). Thermal structure of island arcs. Bull. Geol. Soc. Am. 81, 1665-88.

———— ———— (1971). Origin of paired metamorphic belts and crustal dilation in island arc regions. J. Geophys. Res. 76, 1315-27.

Packham, G.H. (1972). Geological history of the marginal continental areas of the South-west Pacific. In: The Western Pacific: Island Arcs, Marginal Seas, and Geochemistry (ed. P. Coleman). W. Austral. Univ. Press.

———— and Falvey, D.A. (1971). An hypothesis for the formation of marginal seas in the western Pacific. Tectonophysics 11, 79-109.

Page, R.W. (1971). The geochronology of igneous rocks in the New Guinea region. Unpub. Ph.D. thesis, A.N.U.

———— and McDougall, I. (1970). Potassium-argon dating of the Tertiary f_{1-2} stage in New Guinea and its bearing on the geological time-scale. Am. J. Sci. 269, 321-342.

Pakiser, L.C. and Robinson, R. (1966). Composition and evolution of the continental crust as suggested by seismic observations. Tectonophysics 3, 547-557.

Parker, R.L. and Fleischer, M. (1968). Geochemistry of niobium and tantalum. U.S. Geol. Surv. Prof. Pr. 612.

Peacock, M.A. (1931). Classification of igneous rocks. J. Geol. 39, 54-67.

Pearce, J.A. and Cann, J.R. (1971). Ophiolite origin investigated by discriminant analysis using Ti, Zr, and Y. Earth Planet. Sci. Letters 12, 339-349.

Peterman, Z.E., Carmichael, I.S.E. and Smith, A.L. (1970b). Sr 87/Sr86 ratios of Quaternary lavas of the Cascade Range, Northern California. Bull. Geol. Soc. Am. 81, 311-318.

Peterman, Z.E., Doe, B.R. and Prostka, H.F. (1970). Lead and strontium isotopes in rocks of the Absaroka volcanic field, Wyoming. Contr. Mineral. and Petrol. 27, 121-130.

——— and Hedge, C.E. (1971). Related strontium isotopic and chemical variations in oceanic basalts. Bull. Geol. Soc. Am. 82, 493-500.

——— , Lowder, G.G. and Carmichael, I.S.E. (1970a). Sr87/Sr86 ratios of the Talasea series, New Britain, Territory of New Guinea. Bull. Geol. Soc. Am. 81, 39-40.

Phillips, K.A. (1964). Geological Map of Fiji. Fiji Geol. Surv. Ann. Report for 1964.

Philpotts, J.A. (1970). Redox estimation from a calculation of Eu2+ and Eu3+ concentrations in natural phases. Earth Planet. Sci. Letters 9, 257-268.

——— , Martin, W. and Schnetzler, C.C. (1971). Geochemical aspects of some Japanese lavas. Earth Planet. Sci. Letters 12, 89-96.

——— and Schnetzler, C.C. (1970). Phenocryst-matrix partition coefficients for K, Rb, Sr, and Ba, with applications to anorthosite and basalt genesis. Geochim. Cosmochim. Acta 34, 307-322.

——— ——— and Hart, S.R. (1969). Submarine basalts: some K, Rb, Sr, Ba, rare-earth, H_2O and CO_2 data bearing on their alteration, modification by plagioclase, and possible source materials. Earth Planet. Sci. Letters 7, 293-299.

Pitman, W.C.III, Herron, E.M. and Heirtzler, J.R. (1968). Magnetic anomalies in the Pacific and sea floor spreading. J. Geophys. Res. 73, 2069-85.

Pushkar, P. (1968). Strontium isotope ratios in volcanic rocks of three island arc areas. J. Geophys. Res. 73, 2701-13.

——— and Peterson, M.N.A. (1967). Strontium isotopic analyses of three marine phillipsites from the Pacific Ocean. Earth Planet. Sci. Letters 2, 349-350.

Renard, A. (1889). Report on the rock specimens collected on Oceanic Islands during the Voyage of H.M.S. Challenger during the years 1873-76. Physics and Chemistry section, II(4), chapter 11, pp. 149-152.

Rhodes, J.M. and Ridley, W.I. (1971). Trace element abundances in some Andean andesites. Abs. Geol. Soc. Am. Ann. Mtgs. 3, 681-682.

Richards, J.R., Cooper, J.A., Webb, A.W. and Coleman, P.J. (1966), K-Ar measurements of the age of basal schists in the British Solomon Islands. Nature 211, 1251-52.

Rickard, M.J. (1963). The geology of Mbalevuto area. Bull Geol. Surv. Fiji, 11.

——— (1966). Reconnaissance geology of Vanua Levu. Mem. Geol. Surv. Fiji, 2.

——— (1969). Relief of curvature on expansion - a possible mechanism of geosynclinal formation and orogenesis. Tectonophysics 8, 129-144.

Rickard, M.J. (1970). The geology of north-eastern Vanua Levu. Bull. Geol. Surv. Fiji, 14.

———— (in press). Australasia tectonic developments. Encyclopedia Earth Sciences.

Ringwood, A.E. (1966). Mineralogy of the mantle. In: Advances in Earth Science (ed. P.M. Hurley), pp. 357-99. M.I.T. Press.

———— (1969). Composition of the crust and upper mantle. In: The Earth's Crust and Upper Mantle. A.G.U. Mon. 13, 1-17.

———— (1970). Phase transformations and the constitution of the mantle. Phys. Earth Planet. Interiors 3, 109-155.

———— (1972). Phase transformations and mantle dynamics. Earth Planet. Sci. Letters 14, 233-241.

Rittman, A. (1953). Magmatic character and tectonic position of the Indonesian volcanoes. Bull. Volcanol. 14, 45-58.

Robertson, E.I. (1965). Gravity base stations in the south-west Pacific Ocean. N.Z. Jl. Geol. Geophys. 8, 424-439.

———— (1967). Bouguer anomaly map of Viti Levu, Fiji. N.Z. Jl. Geol. Geophys. 10, 1309-12.

Robertson, J.K. and Wyllie, P.J. (1971). Rock-water systems, with special reference to the water-deficient region. Am. J. Sci. 271, 252-277.

Rodda, P. (1967). Outline of the geology of Viti Levu. N.Z. Jl. Geol. and Geophys. 10, 1260-73.

———— (1969). Analyses of rocks from Fiji. Geol. Surv. Dept., Fiji.

———— and Band, R.B. (1966). Geology of Viti Levu. Fiji Geol. Surv. Ann. Report for 1966, pp. 8-16.

———— , Snelling, N.J. and Rex, D.C. (1967). Radiometric age data on rocks from Viti Levu, Fiji. N.Z. Jl. Geol. Geophys. 10, 1248-59.

Rogers, J.J.W. and Donnelly, T.W. (1966). Radiometric evidence for the origin of eugeosynclinal materials. Tulane Studies in Geology 4, 133-138.

Ronov, A.B. and Yaroshevsky, A.A. (1969). Chemical composition of the Earth's crust. In: The Earth's Crust and Upper Mantle, A.G.U. Mon. 13, 37-57.

Rosholt, J.N., Doe, B.R. and Tatsumoto, N. (1966). Evolution of the isotopic composition of uranium and thorium in soil profiles. Bull. Geol. Soc. Am. 77, 987-1004.

———— , Garner, E.L. and Shields, W.R. (1964). Fractionation of uranium isotopes and daughter products in weathered granites and uranium-bearing sandstone, Wind River basin region, Wyoming. U.S. Geol. Surv. Prof. Pr. 501-B, 84-87.

Ruxton, B.P. (1966). A Pleistocene to Recent rhyodacite-trachybasalt-basaltic latite volcanic association in northeast Papua. Bull. Volcanol. 29, 347-374.

Santo, T. (1970). Regional study on the characteristic seismicity of the world. Part III: New Hebrides region. Bull. Earthquake Res. Inst. 48, 1-18.

Sass, J.H. (1971). Tectonic and geochemical implications of heat flow data. Comments on Earth Sci.: Geophysics 1, 87-94.

———— and Munroe, R.J. (1970). Heat flow from deep boreholes on two island arcs. J. Geophys. Res. 75, 4387-95.

Schilling, J.G. (1971). Sea-floor evolution: rare-earth evidence. Phil. Trans. Roy. Soc. Lond. A 268, 663-706.

Schmidt, R.G. (1957). Petrology of the volcanic rocks, Saipan, Mariana Islands. U.S. Geol. Surv. Prof. Pr. 280, 127-176.

Schnetzler, C.C. and Philpotts, J.A. (1970). Partition coefficients of rare-earth elements between igneous matrix material and rock-forming mineral phenocrysts - II. Geochim. Cosmochim. Acta 34, 331-340.

Scholz, C.H., Barazangi, M. and Sbar, M.L. (1971). Late Cenozoic evolution of the Great Basin, Western United States, as an ensialic inter-arc basin. Bull. Geol. Soc. Am. 82, 2979-3011.

Sclater, J.G., Anderson, R.N. and Bell, M.L. (1971). Evelation of ridges and evolution of the Central Eastern Pacific. J. Geophys. Res. 76, 7888-7915.

———— , Hawkins, J.W., Mammerickx, J. and Chase, C.G. (1972). Crustal extension between the Tonga and Lau Ridges: petrologic and geophysical evidence. Bull. Geol. Soc. Am. 83, 505-518.

———— and Menard, H.W. (1967). Topography and heat flow of the Fijian Plateau. Nature 216, 991-993.

Scott, R.B., Nesbitt, R.W., Dasch, E.J. and Armstrong, R.L. (1971). A strontium isotope evolution model for Cenozoic magma genesis, Eastern Great Basin, U.S.A. Bull. Volcanol. 35, 1-26.

Seeley, J.B. and Searle, E.J. (1970). Geology of the Rakiraki district, Viti Levu, Fiji. N.Z. Jl. Geol. Geophys. 13, 52-71.

Shaw, D.M. (1970). Trace element fractionation during anatexis. Geochim. Cosmochim. Acta 34, 237-242.

Shido, F., Miyashiro, A. and Ewing, M. (1971). Crystallization of abyssal tholeiites. Contr. Mineral. and Petrol. 31, 251-266.

Shor, G.G.Jr., Kirk, H.K. and Menard, H.W. (1971). Crustal structure of the Melanesian area. J. Geophys. Res. 76, 2562-86.

———— , Menard, H.W., and Raitt, R.W. (1970). Structure of the Pacific Basin. In: The Sea (ed. A.E. Maxwell), Vol. 4, Part II, pp. 3-28.

Smith, A.L. and Carmichael, I.S.E. (1968). Quaternary lavas from the southern Cascades, Western U.S.A. Contr. Mineral. and Petrol. 19, 212-238.

Smith, I.E. (1970). Late Cainozoic volcanism in Eastern Papua. Austral. B.M.R. Record 1970/72.

Smith, J.G. and Dickinson, W.R. (1965). A geological reconnaissance of the southern Yasawa Islands. Fiji Geol. Surv. Note 125.

Smith, R.E. (1968). Redistribution of major elements in the alteration of some basic lavas during burial metamorphism. J. Petrology 9, 191-219.

Solomon, S. and Biehler, S. (1969). Crustal structure from gravity anomalies in the southwest Pacific. J. Geophys. Res. 74, 6696-6701.

Stanton, R.L. and Bell, J.D. (1969). Volcanic and associated rocks of the New Georgia Group, British Solomon Islands Protectorate. Overseas Geol. Min. Resources 10, 113-145.

Stearns, H.T. (1971). Geologic setting of an Eocene fossil deposit on Eua Island, Tonga. Bull. Geol. Soc. Am. 82, 2541-52.

Stice, G.D. (1968). Petrography of the Manu'a Islands, Samoa. Contr. Mineral. and Petrol. 19, 343-357.

Stillwell, F.L. (1949). Occurrence of telluride minerals at Vatukoula, Fiji. Proc. Aust. Inst. Min. Met. No. 154-155, 3-27.

Stipp, J.J. (1968). The geochronology and petrogenesis of the Cenozoic volcanic rocks of North Island, New Zealand. Unpub. Ph.D. thesis, A.N.U.

Suess, E. (1904). The Face of the Earth, vol. 4. Oxford: Clarendon Press.

Sugimura, A. (1961). Regional variation of the K_2O/Na_2O ratios of volcanic rocks in Japan and environs. J. Geol. Soc. Japan 67, 292-300.

———— (1968). Spatial relations of basaltic magmas in island arcs. In: Basalts (eds. H.H. Hess and A. Poldevaart), 2, 537-571. Interscience Pub.

———— (1972). Multiple correlation between composition of volcanic rocks and depth of earthquake foci. In: The Western Pacific: Island Arcs, Marginal Seas, and Geochemistry (ed. P. Coleman). W. Austral. Univ. Press.

———— , Matsuda, T., Chinzei, K. and Nakamura, K. (1963). Quantitative distribution of late Cenozoic volcanic materials in Japan. Bull. Volcanol. 26, 125-140.

Sutton, G.H., Maynard, L. and Hussong, D. (1970). Crustal structure: the widespread occurrence of the 7.4-7.6 km/sec basal crustal layer in the Pacific Ocean Basin. H.I.G. Rep. 70-27, pp. 53-63.

Switzer, G. and Melson, W.G. (1969). Partially melted kyanite eclogite from the Roberts Victor Mine, South Africa. Smithsonian Contr. Earth Sciences No. 1.

Sykes, L.R., Isacks, B.L. and Oliver, J. (1968). Spatial distribution of deep and shallow earthquakes of small magnitude in the Fiji-Tonga region. Bull. Seism. Soc. Am. 59, 1093-1114.

Tatsumoto, M. (1969). Lead isotopes in volcanic rocks and possible ocean-floor thrusting beneath island arcs. Earth Planet. Sci. Letters 6, 369-376.

———— and Knight, R.J. (1969). Isotopic composition of lead in volcanic rocks from central Honshu - with regard to basalt genesis. Geochem. J. 3, 53-86.

Taylor, S.R. (1964). Abundance of chemical elements in the continental crust: a new table. Geochim. Cosmochim. Acta 28, 1273-85.

———— (1965). Geochemical analysis by spark source mass spectrography. Geochim. Cosmochim. Acta 29, 1243-61.

Taylor, S.R. (1967). The origin and growth of continents. Tectonophysics 4, 17-34.

———— (1969). Trace element chemistry of andesites and associated calc-alkaline rocks. In: Proc. Andesite Conf. (ed. A.R. McBirney), Dept. Geol. Min. Res. Oregon Bull. 65, 43-64.

———— (1971). Geochemical application of spark source mass spectrometry II: photoplate data processing. Geochim. Cosmochim. Acta 35, 1187-96.

———— , Capp, A.C., Graham, A.L. and Blake, D.H. (1969). Trace element abundances in andesites II. Saipan, Bougainville and Fiji. Contr. Mineral. and Petrol. 23, 1-26.

———— , Kaye, M., White, A.J.R., Duncan, A.R. and Ewart, A. (1969). Genetic significance of Co, Cr, Ni, Sc, and V in andesites. Geochim. Cosmochim. Acta 33, 275-286.

———— and White, A.J.R. (1966). Trace element abundances in andesites. Bull. Volcanol. 29, 177-194.

Thompson, G. (1968). Analyses of B, Ga, Rb, and K in two deep-sea sediment cores. Marine Geol. 6, 463-477.

———— and Melson, W.G. (1970). Boron contents of serpentinites and metabasalts in the oceanic crust: implications for the boron cycle in the oceans. Earth Planet. Sci. Letters 8, 61-65.

Thomson, A.A. and Evison, F.F. (1962). Thickness of the Earth's crust in New Zealand. N.Z. Jl. Geol. Geophys. 5, 29-45.

Tilling, R.I., Gottfried, D. and Dodge F.C.W. (1970). Radiogenic heat production of contrasting magma series: bearing on interpretation of heat flow. Bull. Geol. Soc. Am. 81, 1447-62.

Toksöz, M.N., Minear, J.W and Julian, B.R. (1971). Temperature field and geophysical effects of a downgoing slab. J. Geophys. Res. 76, 1113-38.

Tomita, T. (1935). On the chemical compositions of the Cenozoic alkaline suite of the circum-Japan-Sea region. Shanghai Sci. Inst. Jour. 1, 227-306.

Turekian, K.K. and Wedepohl, K.H. (1961). Distribution of the elements in some major units of the Earth's crust. Bull. Geol. Soc. Am. 72, 175-192.

Utsu, T. (1971). Seismological evidence for anomalous structure of island arcs with special reference to the Japanese region. Rev. Geophys. Space Phys. 9, 839-890.

Uyeda, S. and Horai, K. (1964). Terrestrial heat flow in Japan. J. Geophys. Res. 69, 2121-41.

van Bemmelen, R.W. (1966). On mega-undations: a new model for the Earth's evolution. Tectonophysics 3, 83-127.

van der Linden, W.J.M. (1967). Structural relationships in the Tasman Sea and south-west Pacific Ocean. N.Z. Jl. Geol. Geophys. 10, 1280-1301.

Varne, R. (1970). Hornblende lherzolite from the upper mantle. Contr. Mineral. and Petrol. 27, 45-51.

Varne, R. and Graham, A.L. (1971). Rare earth abundances in hornblende and clinopyroxene of a hornblende lherzolite xenolith: implications for upper mantle fractionation processes. Earth Planet. Sci. Letters 13, 11-18.

Veizer, J. (1971). Chemical and strontium isotopic evolution of sedimentary carbonate rocks in geologic history. Unpub. Ph.D. thesis, A.N.U.

Velde, B., Hervé, F. and Kornprobst, J. (1970). The eclogite-amphibolite transition at 650°C and 6.5 kbar pressure, as exemplified by basic rocks of the Uzerche area, Central France. Am. Miner. 55, 953-974.

Verhoogen, J. (1962). Distribution of titanium between silicates and oxides in igneous rocks. Am. J. Sci. 260, 211-220.

Vine, F.J. and Hess, H.H. (1971). Sea floor spreading. In: The Sea (ed. A. Maxwell), Vol. 4, Part 2, pp. 587-622.

Vinogradov, A.P. (1962). Average content of chemical elements in main types of igneous rocks of the earth's crust. Geokhimiya (7), 555-571.

Vlasov, K.A. (ed.) (1966). Geochemistry and Mineralogy of Rare Elements and Genetic Types of their Deposits. 3 volumes. Jerusalem: Israel Program for Scientific Translations.

Vlodavetz, V.I. and Piip, B.I. (1959). Catalogue of the active volcanoes of the world. Part 8, Kamchatka and continental areas of Asia. Int. Assoc. Volcanology.

Vogt, P.R. and Conolly, J.R. (1971). Tasmantid guyots, the age of the Tasman Basin, and motion between the Australian plate and the mantle. Bull. Geol. Soc. Am. 82, 2577-84.

Washington, H.S. (1917). Chemical analyses of igneous rocks. U.S. Geol. Surv. Prof. Pr. 99.

Watkins, N.D., Gunn, B.M. and Coy-yll, R. (1970). Major and trace element variations during the initial cooling of an Icelandic lava. Am. J. Sci. 268, 24-49.

Wedepohl, K.H. (1968). Chemical fractionation in the sedimentary environment. In: Origin and Distribution of the Elements (ed. L.H. Ahrens), pp. 999-1016. Pergamon Press.

———— (1969). Composition and abundance of common sedimentary rocks. In: Handbook of Geochemistry (ed. K.H. Wedepohl), Vol. I, 250-271. Springer-Verlag.

Wellman, P., McElhinney, M.W. and McDougall, I. (1969). On the polar-wander path for Australia during the Cenozoic. Geophys. J. R. Astr. Soc. 18, 371-395.

White, A.J.R. (unpublished manuscript). Petrography of some stone adzes from Tongatapu, Tonga Group.

Willmott, W.F. (1972). Late Cainozoic shoshonitic lavas from Torres Strait, Papua and Queensland. Nature 235, 33-35.

Wilson, J.T. (1965). A new class of faults and their bearing on continental drift. Nature 207, 343-347.

Wolf, K.H., Chilingar, G.V. and Beales, F.W. (1967). Elemental composition of carbonate skeletons, minerals, and sediments. In: Carbonate Rocks (eds. G.V. Chilingar, H.F. Bissell, and R.W. Fairbridge), Vol. II, pp. 23-150. Elsevier Pub. Co.

Woollard, G.P. and Sutton, G.H. (1970). Geophysical and geological investigations in areas of anomalous gravity. H.I.G. Rep. 70-27, pp. 8-47.

Woolnough, W.G. (1903). The continental origin of Fiji. Proc. Linn. Soc. N.S.W. 28, 457-540.

Wright, T.L. and Doherty, P.C. (1970). A linear programming and least squares computer method for solving petrologic mixing problems. Bull. Geol. Soc. Am. 81, 1995-2008.

Wyllie, P.J. (1971a). The Dynamic Earth. John Wiley and Sons, Inc. 416 pp.

———— (1971b). Role of water in magma generation and initiation of diapiric uprise in the mantle. J. Geophys. Res. 76, 1328-38.

Yagi, K., Kawano, Y. and Aoki, K. (1963). Types of Quaternary volcanic activity in Northeastern Japan. Bull. Volcanol. 26, 223-235.

Yajima, T., Higuchi, H., Banno, S. and Nagasawa, H. (1968). Differentiation of rocks in Izu-Hakone region; a study of rare earth patterns. Chikyukagaku 2, 24-25.

Yoder, H.S.Jr. (1969). Calcalkalic andesites: experimental data bearing on the origin of their assumed characteristics. In: Proc. Andesite Conf. (ed. A.R. McBirney), Dept. Geol. Min. Res. Oregon Bull. 65, 77-90.

———— and Kushiro, I. (1969). Melting of a hydrous phase: phlogopite. Am. J. Sci. 267-A, 558-582.

———— and Tilley, C.E. (1962). Origin of basalt magmas: an experimental study of natural and synthetic rock systems. J. Petrology 3, 342-532.

———— and Upton, B.G.J. (1971). Diopside-sanidine-H_2O at 5 and 10 kb. Carnegie Inst. Year Bk. 70, 108-112.

Appendix 1. Analytical Techniques

A1.1 Introduction

Samples were processed in batches of 30, 130, and 25 at approximately one year intervals and are so designated by their 68-, 69-, and 71- prefixes in Appendix 2. Some variation in technique and operating conditions was unavoidable but minor, and samples from previous batches were re-analyzed to verify comparability. Analytical variability is certainly less than that observable in Table 1.2.

These techniques were developed and their requisite machinery maintained largely by other people to whom I am grateful for permission to use them and for instruction. I was solely in charge of the gamma-ray spectrometer throughout most of my study, the XRF spectrometer for most of my major element analyses, and a wet chemistry laboratory for Sr separations. I, in turn, instructed others in these and spectrographic techniques and shared responsibility for setting up and calibrating other equipment and procedures.

A1.2 Sample preparation

All weathered surfaces were removed with a jaw splitter and 1-2 kg of sample processed. It was crushed to at least <120 mesh powder (Kleeman, 1967) by passes through a steel jaw crusher (to $\frac{1}{4}$" size), an agate cone grinder (to <20 mesh), and either an agate siebtechnik or ball mill. Care was taken to prevent cross-contamination or metal contamination from grinding. Standard quartering techniques and/or a Perspex sample cutter were used to split samples after cone grinding to a 60 g amount which was drawn from thereafter for analysis; separate aliquots were taken for gamma-ray spectrometry and K-Ar work. The 60 g analysis aliquots were stored in sealed glass jars and dispensed, after shaking of the jar, with weighing paper rather than metal spatulae. Further grinding in an agate mortar or ball mill was necessary for mass spectrography.

A1.3 Major element analysis

A. X-ray fluorescence

Concentrations of Si, Ti, Al, Fe, Mn, Mg, Ca, K, and P were measured on a modified Phillips PW 1220 semi-automatic X-ray spectrometer operated under the conditions listed in Table A1, and following the technique of Norrish and Hutton (1969). In this method rough weight per cent oxide concentrations are determined relative to USGS standard rocks.

Table A1. XRF operating conditions and count rates.

Element	Peak (2θ)	Analyzing Crystal	Tube	kV	mA	Flow Detector	Lower Level	Window	Background (as %)	CPS/% or ppm	Standard Analysis Time (sec)
			A. MAJOR ELEMENTS								
Mg	137.76	ADP	Cr	44	36	1710.	270.	190.	0.385	4.7	200
Al	144.8	PE	"	"	"	1647.	200.	200.	.144	70.	100
Si	108.98	PE	"	"	"	"	"	"	.400	73.	"
P	140.86	GE	"	"	"	1617.	"	"	.057	176.	"
K	136.74	LiF	"	"	8	1567.	"	"	.021	2113.	40
Ca	113.11	"	"	"	9	"	"	"	.082	2375.	"
Ti	86.12	"	"	"	10	"	"	"	.118	3417.	"
Mn	62.86	"	W	55	36	1580.	"	"	.110	2693.	"
Fe	57.39	"	"	"	"	"	"	"	.136	886.	"
			B. TRACE ELEMENTS								
Rb	26.57	LiF(200)	Mo	90	20	880.	30.	100.	-	2.5*	200
Sr	25.15	"	"	"	"	"	"	"	-	3.5	"
Th	27.48	"	"	"	"	"	"	"	-	1.3	"
U	22.16	"	"	"	"	"	"	"	-	1.3	"
Pb	28.27	"	"	"	"	"	"	"	-	0.91	"
Y	33.86	LiF(220)	"	"	"	"	"	"	-	2.0	"
Zr	22.56	"	"	"	"	"	"	"	-	1.3	"

* Dependent on mass absorption. Values given are averages for BCR-1, W-1, AGV-1, GSP-1, and G-2.

Table A.2. Major element concentrations measured for interlaboratory standard rocks. Numbers in parentheses are averages of values obtained by C.S.I.R.O., Adelaide, and Department of Geology, A.N.U., 1970. Na_2O values are from E. Kiss.

	AGU-1		BCR-1		GSP-1	
SiO_2	58.36	(58.99)	53.92	(53.92)	67.24	(67.25)
σ	.24		.18			
c.v.	.42		.34			
TiO_2	1.06	(1.05)	2.25	(2.25)	0.66	(0.66)
σ	.0049		.0050			
c.v.	.42		.21			
Al_2O_3	17.04	(17.00)	13.36	(13.47)	15.08	(15.00)
σ	.087		.064			
c.v.	.50		.48			
Fe_2O_3	6.80	(6.74)	13.39	(13.41)		
σ	.028		.042			
c.v.	.40		.32			
MnO	0.10	(0.10)	0.18	(0.19)		
σ	.0016		.0013			
c.v.	.81		.44			
MgO	1.52	(1.53)	3.48	(3.44)		
σ	.052		.070			
c.v.	2.73		1.84			
CaO	4.93	(4.90)	6.98	(6.98)	2.03	(1.99)
σ	.011		.013			
c.v.	.21		.18			
K_2O	2.90	(2.90)	1.74	(1.70)	5.52	(5.48)
σ	.0070		.016			
c.v.	.24		.93			
P_2O_5	0.48	(0.50)	0.36	(0.37)	0.27	(0.28)
σ	.0071		.014			
c.v.	1.26		3.32			
Na_2O	4.34	(4.33)				
L.O.I.	1.76	(1.69)				
σ	.051					
c.v.	2.9					
TOTAL	99.80					

Table A.3. Lower limits of detection (LLD) for trace element analyses.*

Element	Method	LLD** (ppm)
B	E	10.
Ba	E	3.
Ce	S	0.1
Co	E	5.
Cr	E	1.
Cu	E	0.5
Dy	S	0.1
Er	S	0.1
Eu	S	0.1
Gd	S	0.1
Hf	S	0.1
Ho	S	0.1
La	S	0.1
Nd	S	0.1
Ni	E	5.
Pb	X	1.0*
Pr	S	0.1
Rb	X	0.6*
Sc	E	2.
Sm	S	0.1
Sn	S	0.1
Sr	X	0.5*
Tb	S	0.1
Th	G	0.09*
Tm	S	0.1
U	G	0.03*
V	E	5.
Y	X	0.7*
Yb	S	0.1
Zr	X	1.3*

Notes

*Based on counting statistics where LLD (95% confidence) = $3/M \sqrt{Cb/Tb}$ and M = cps/ppm

Cb = cps background

Tb = time counting background when Tb $\cong T_{peak}$.

**LLD varies with mass absorption; values given are averages for rocks mentioned in Table A1.1.

Table A.4. Measured trace element concentrations for interlaboratory standard rocks, lunar fines, and some Fijian samples.*

Element	Analytical Method	G-1	W-1	AGV-1	GSP-1	G-2	BCR-1	S-1	PCC-1	DTS-1	BHVO-1	NBS 70a	Apollo 10084	Apollo 12070	814	816	825	874
B	E	-	10.	-	-	-	-	74.	-	-	-	-	-	-	-	-	-	-
Ba	E	1050.	160.	1200.	1250.	1850.	740.	320.	-	-	132.	-	-	-	287.	-	297.	541.
σ		-	-	-	-	-	-	-	-	-	9.	-	-	-	14.	-	21.	35.
c.v.		-	-	-	-	-	-	-	-	-	7.0	-	-	-	4.7	-	7.1	6.5
N		-	-	-	-	-	-	-	-	-	16	-	-	-	8	-	16	24.
Ce	S	170.	23.	-	-	-	-	-	-	-	33.	-	47.	84.	-	-	-	22.
σ		-	-	-	-	-	-	-	-	-	2.6	-	-	-	-	-	-	1.1
c.v.		-	-	-	-	-	-	-	-	-	7.8	-	-	-	-	-	-	5.1
N		-	-	-	-	-	-	-	-	-	16	-	-	-	-	-	-	10
Co	E	2.3	47.	16.	7.0	-	40.	19.	120.	143.	-	-	-	-	-	-	34.	35.
σ		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	1.3	3.1
c.v.		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	3.7	8.8
N		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	9	5
Cr	E	20.	115.	10.	12.	8.0	14.	55.	-	-	-	-	-	-	-	-	30.	69.
σ		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	2.5	4.9
c.v.		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	8.3	7.1
N		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	8	5
Cu	E	13.	115.	60.	32.	11.	22.	20.	9.2	7.0	-	-	-	-	-	-	58.	-
σ		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	3.8	-
c.v.		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	6.5	-
N		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	9	
Dy	S	-	3.9	3.5	-	-	-	-	-	-	4.7	-	20.	23.	-	-	-	3.0
σ		-	-	-	-	-	-	-	-	-	0.6	-	-	-	-	-	-	0.2
c.v.		-	-	-	-	-	-	-	-	-	12.0	-	-	-	-	-	-	5.8
N		-	-	-	-	-	-	-	-	-	16	-	-	-	-	-	-	16
Er	S	-	2.3	2.0	-	-	-	-	-	-	1.8	-	12.	15.	-	-	-	1.4
σ		-	-	-	-	-	-	-	-	-	0.3	-	-	-	-	-	-	0.1
c.v.		-	-	-	-	-	-	-	-	-	15.5	-	-	-	-	-	-	4.2
N		-	-	-	-	-	-	-	-	-	16	-	-	-	-	-	-	16
Eu	S	1.2	1.1	1.6	-	-	-	-	-	-	1.7	-	1.7	1.9	-	-	-	1.0
σ		-	-	-	-	-	-	-	-	-	0.2	-	-	-	-	-	-	0.1
c.v.		-	-	-	-	-	-	-	-	-	11.2	-	-	-	-	-	-	6.1
N		-	-	-	-	-	-	-	-	-	16	-	-	-	-	-	-	16
Gd	S	-	4.0	6.1	-	-	-	-	-	-	5.6	-	15.	25.	-	-	-	3.4
σ		-	-	-	-	-	-	-	-	-	0.2	-	-	-	-	-	-	0.2
c.v.		-	-	-	-	-	-	-	-	-	3.6	-	-	-	-	-	-	7.2
N		-	-	-	-	-	-	-	-	-	16	-	-	-	-	-	-	16
Hf	S	-	2.0	5.0	-	-	-	-	-	-	4.0	-	10.	12.	-	-	-	1.6
σ		-	-	-	-	-	-	-	-	-	0.6	-	-	-	-	-	-	0.1
c.v.		-	-	-	-	-	-	-	-	-	14.6	-	-	-	-	-	-	7.0
N		-	-	-	-	-	-	-	-	-	16	-	-	-	-	-	-	24
Ho	S	0.5	1.0	0.8	-	-	-	-	-	-	0.90	-	4.5	6.0	-	-	-	0.53
σ		-	-	-	-	-	-	-	-	-	0.14	-	-	-	-	-	-	0.02
c.v.		-	-	-	-	-	-	-	-	-	15.5	-	-	-	-	-	-	4.9
N		-	-	-	-	-	-	-	-	-	16	-	-	-	-	-	-	8

Table A.4. continued

Element	Analytical Method	G-1	W-1	AGV-1	GSP-1	G-2	BCR-1	S-1	PCC-1	DTS-1	BHVO-1	NBS 70a	Apollo 10084	Apollo 12070	814	816	825	874
La	S	100.	10.	33.	-	-	-	-	-	-	18.	-	16.5	32.	-	-	-	10.5
σ		-	-	-	-	-	-	-	-	-	1.4	-	-	-	-	-	-	0.7
c.v.		-	-	-	-	-	-	-	-	-	8.0	-	-	-	-	-	-	6.3
N		-	-	-	-	-	-	-	-	-	16	-	-	-	-	-	-	9
Nd	S	56.	15.	30.	-	-	-	-	-	-	22.	-	39.	54.	-	-	-	14.
σ		-	-	-	-	-	-	-	-	-	2.4	-	-	-	-	-	-	0.7
c.v.		-	-	-	-	-	-	-	-	-	11.1	-	-	-	-	-	-	4.8
N		-	-	-	-	-	-	-	-	-	16	-	-	-	-	-	-	14
Ni	E	-	75.	16.	8.0	4.5	11.	37.	-	-	-	-	-	-	40.	-	28.	-
σ		-	-	-	-	-	-	-	-	-	-	-	-	-	2.6	-	1.0	-
c.v.		-	-	-	-	-	-	-	-	-	-	-	-	-	6.%	-	3.8	-
N		-	-	-	-	-	-	-	-	-	-	-	-	-	5	-	9	-
Pb	X	-	6.8	36.	54.	31.	13.	-	-	-	-	-	-	-	-	-	-	-
σ		-	1.1	-	-	-	1.4	-	-	-	-	-	-	-	-	-	-	-
c.v.		-	17.	-	-	-	11.	-	-	-	-	-	-	-	-	-	-	-
N		-	15	-	-	-	20	-	-	-	-	-	-	-	-	-	-	-
Pr	S	18.	3.6	10.7	-	-	-	-	-	-	5.5	-	7.0	-	-	-	-	3.5
σ		-	-	-	-	-	-	-	-	-	0.6	-	-	-	-	-	-	0.2
c.v.		-	-	-	-	-	-	-	-	-	11.7	-	-	-	-	-	-	5.9
N		-	-	-	-	-	-	-	-	-	16	-	-	-	-	-	-	9
Rb	X	-	22.	69.	253.	170.	48.	-	-	-	-	521.	-	-	-	-	-	-
σ		-	0.8	-	-	-	1.2	-	-	-	-	-	-	-	-	-	-	-
c.v.		-	3.7	-	-	-	2.5	-	-	-	-	-	-	-	-	-	-	-
N		-	15	-	-	-	20	-	-	-	-	-	-	-	-	-	-	-
Sc	E	2.8	34.	11.5	6.6	3.4	32.	-	9.0	-	-	-	-	-	-	-	26.	-
σ		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.9	-
c.v.		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	3.4	-
N		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	8	-
Sm	S	8.3	3.5	6.0	-	-	-	-	-	-	5.3	-	13.	19.	-	-	-	3.3
σ		-	-	-	-	-	-	-	-	-	0.6	-	-	-	-	-	-	0.2
c.v.		-	-	-	-	-	-	-	-	-	11.7	-	-	-	-	-	-	6.3
N		-	-	-	-	-	-	-	-	-	16	-	-	-	-	-	-	16
Sn	S	3.0	1.1	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.26
σ		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.02
c.v.		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	5.6
N		-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	5
Sr	X	-	190.	668.	238.	484.	332.	-	-	-	-	-	-	-	-	-	-	-
σ		-	3.1	-	-	-	5.4	-	-	-	-	-	-	-	-	-	-	-
c.v.		-	1.6	-	-	-	1.6	-	-	-	-	-	-	-	-	-	-	-
N		-	15	-	-	-	20	-	-	-	-	-	-	-	-	-	-	-
Tb	S	0.6	0.7	0.7	-	-	-	-	-	-	0.75	-	3.2	3.2	-	-	-	0.50
σ		-	-	-	-	-	-	-	-	-	0.13	-	-	-	-	-	-	0.03
c.v.		-	-	-	-	-	-	-	-	-	17.3	-	-	-	-	-	-	6.7
N		-	-	-	-	-	-	-	-	-	16	-	-	-	-	-	-	7

Table A.4. continued

Element	Analytical Method	G-1	W-1	AGV-1	GSP-1	G-2	BCR-1	S-1	PCC-1	DTS-1	BHVO-1	NBS 70a	Apollo 10084	Apollo 12070	814	816	825	874
Th	G	-	-	6.3	106.	25.5	6.1	-	-	-	-	-	-	-	-	0.49	-	-
σ		-	-	-	-	-	0.2	-	-	-	-	-	-	-	-	0.12	-	-
c.v.		-	-	-	-	-	7.7	-	-	-	-	-	-	-	-	24.	-	-
N		-	-	-	-	-	10	-	-	-	-	-	-	-	-	11	-	-
U	G	-	-	1.83	2.36	2.17	1.61	-	-	-	-	-	-	-	-	0.18	-	-
σ		-	-	-	-	-	0.04	-	-	-	-	-	-	-	-	0.04	-	-
c.v.		-	-	-	-	-	4.5	-	-	-	-	-	-	-	-	20.	-	-
N		-	-	-	-	-	10	-	-	-	-	-	-	-	-	11	-	-
V	E	14.	245.	115.	52.	35.	380.	88.	31.	-	-	-	-	-	298.	-	265.	-
σ		-	-	-	-	-	-	-	-	-	-	-	-	-	18.	-	14.	-
c.v.		-	-	-	-	-	-	-	-	-	-	-	-	-	6.1	-	5.3	-
N		-	-	-	-	-	-	-	-	-	-	-	-	-	5	-	8	-
Y	X	-	26.	22.	27.	11.	42.	-	-	-	-	-	-	-	-	-	-	-
σ		-	0.7	-	-	-	1.1	-	-	-	-	-	-	-	-	-	-	-
c.v.		-	2.8	-	-	-	2.7	-	-	-	-	-	-	-	-	-	-	-
N		-	18	-	-	-	18	-	-	-	-	-	-	-	-	-	-	-
Yb	S	0.95	2.0	-	-	-	-	-	-	-	1.6	-	12.	12.	-	-	-	1.4
σ		-	-	-	-	-	-	-	-	-	0.3	-	-	-	-	-	-	0.1
c.v.		-	-	-	-	-	-	-	-	-	16.1	-	-	-	-	-	-	4.8
N		-	-	-	-	-	-	-	-	-	16	-	-	-	-	-	-	20
Zr	X	-	89.	220.	-	302.	176.	-	-	-	-	-	-	-	-	-	-	-
σ		-	4.	10.	-	4.	6.	-	-	-	-	-	-	-	-	-	-	-
c.v.		-	4.5	4.5	-	1.3	3.4	-	-	-	-	-	-	-	-	-	-	-
N		-	18	18	-	8	18	-	-	-	-	-	-	-	-	-	-	-

*Analytical precision varies with each measurement and the statistical information is given only as a representative guide to what uncertainties should be expected. N denotes the number of replications used in computations except for Fiji sample 69-874 where it refers to the number of intensity ratios obtained during an analysis involving the exposure of two plates.

Methods are keyed as follows : X = XRF; G = gamma-ray spectrometry; E = emission spectrometry; and S = spark source mass spectrometry.

This provisional analysis together with independently measured Na_2O and loss on ignition values and minus background effects is then used to calculate mass absorption coefficients and, eventually, final concentrations for each element using an iterative computer program written at C.S.I.R.O., Adelaide. No corrections were made for fluorescence effects of other radiations on Mg K_α intensities as these were observed to cancel one another. All samples were analyzed in duplicate and some buttons were cross-checked by analysis in the Department of Geology, A.N.U., by G. Bradley.

Precision and accuracy can be assessed from statistical treatment of replicate measurements of international standards, which is provided in Table A2.

B. Flame photometry

Na concentrations were measured using a modified Perkin Elmer Model 146 flame photometer and the technique of Cooper (1963). A few samples were measured in duplicate and they agreed to within 0.3%. Accuracy was monitored by measuring %K for the same solutions, and values thus obtained agreed with those determined by XRF for the same sample to within 0.019 ±.02 weight per cent.

C. Other

Fe^{+2} was measured for the 68- series using the titration technique of Kiss (1967) who also discusses its reliability. All measurements were in duplicate.

Loss on ignition (LOI) was measured by heating 0.5-1.0 g of sample at 1000°C for one hour in pre-heated quartz crucibles which were monitored carefully to guarantee their constant weight. All determinations were in duplicate or triplicate. Results of replicate analyses of AGV-1 are given in Table A2.

The average total of these oxide and LOI measurements for rocks in my 1970 study was 99.42 ±.42 weight per cent, and 99.68 ±.31 for 1971 analyses.

A1.4 Trace element analysis

A. X-ray fluorescence

Several elements were detected by various XRF methods. All samples were analyzed in duplicate and for sufficient time to obtain 10^4 counts at peak positions, following the general procedure of Norrish and Chappel (1967).

The 68- series samples were analyzed for Rb, Sr, Pb, Y, Zr, and Ba using directly measured mass absorption coefficients (Norrish and Chappel, 1967, pp. 194-6). The mean of two background measurements was subtracted from peak heights, and inter-element and non-linearity corrections applied. Rb and Sr K_α mass absorption coefficients were measured in duplicate and usually agreed to within 2%. The Rb coefficient was also used in calculation of Pb concentrations; the Sr value used for Y and Zr; and an empirical function of the Rb value used for Ba. Comparison of results in Table 1.2 suggests that Y values are 5 to 10 ppm too low in high Rb samples. The 69- series samples were analyzed for Rb, Sr, and Pb, in an identical manner.

The 69- series samples were analyzed for Y and Zr contents, and the 71- series samples for these elements plus Rb, Sr, and Pb by using corrected background intensity as a measure of mass absorption. Each background as well as peak position was separately corrected for all inter-element effects. Because reciprocals of background intensities are, by themselves, unsatisfactory estimates of mass absorption (e.g. Fairbairn and Hurley, 1971), the mean corrected background intensities were multiplied by an empirically determined, wavelength-dependent factor and the result used to correct for mass absorption. Operating conditions used are given in Table A1. Zr values may be about 3% too low when Th contents exceed 3 ppm as Th interference on Zr background was not monitored.

Both methods were calibrated using the concentrations given in Table A4 for USGS standards. Precision, based on replicate analyses, is also indicated in that table, and is generally better than 5% at the level of concentrations encountered. Rb contents less than 10 ppm reflect recycled analyses where $time_{peak}$ = 1600 seconds. Precision for this counting time, based on counting statistics, is about 3% at the 1 ppm Rb level. Most Pb concentrations were at or near detection limits for $time_{peak}$ = 400 seconds. They may be, therefore, of order of magnitude significance only, although there has been good agreement ($\pm$0.2 ppm) between XRF and isotope dilution Pb analyses at the 2 to 5 ppm level in this laboratory.

Accuracy of each method can be assessed from data in Table A4. Measured mass absorption coefficients permitted more accurate analyses than estimates from background intensities, however carefully measured and corrected. Nevertheless, the latter method gave adequate results. Several samples were analyzed by both methods and disagreement rarely exceeded 5%.

B. Gamma-ray Spectrometry

K, U, and Th concentrations were measured using a gamma-ray spectrometric method similar to those described by Heier and Rogers (1963), Adams and Gasparine (1970), and Cherry et al. (1970). The K40 (1.47 MeV), Bi214 (1.76 MeV), and T1208 (2.62 MeV) photopeaks were monitored for K, U, and Th respectively using a 5x4" NaI(T1) analyzing crystal and approximately 30 channels per element of a 400-channel R.I.D.L. Model 24-2 pulse height analyzer. Approximately 500 g aliquots of sample were sealed into round, flat, plastic containers (5" diameter, 1" height) and allowed to equilibrate with Ra226 for three weeks before counting. No absorption corrections were made. Use of these indirect U and Th measurements assumes complete decay chain equilibrium which should be true for unweathered rocks older than 10^6 years. The assumption could be false for some Fijian rocks (Cherdyntsev and Senina, 1970) but data is inadequate for assessment.

Synthetic standards were used for periodic calibrations. In addition to normal calibration procedures (Adams and Gasparine, 1970, pp. 148ff) where allowance is made for the contribution of higher energy gamma rays, high K/U ratios in Fijian shoshonites made it necessary to correct for effects of the "potassium tail" in the 1.6-1.8 MeV portion of typical spectra. The size of this tail was proportional to per cent K and not, therefore, due merely to the 1-2 ppm K in most NaI(T1) crystals. This correction is negligible in rocks of normal K/U but can be 10-15% when K/U $\sim$50,000.

Precision and accuracy can be assessed from data compiled in Table A4 for interlaboratory standards and Fijian rocks of commonly encountered U and Th contents. Accuracy was further monitored by comparing measured K concentrations with those determined by XRF for the same sample with which they agreed to within .0183 $\pm$.015 weight per cent.

Average 1970 backgrounds, count rates, and lower detection limits based on counting statistics alone and my standard 1400 minute counting time are, respectively, as follows: for K, .764%, 46.77 cpm/%, .014%; for U, 4.260 ppm, 3.69 cpm/ppm, 0.12 ppm; and for Th, 9.810 ppm, 1.05 cpm/ppm, 0.34 ppm. (See footnote, Table A3, for calculation of detection limits.) Many Fijian samples have U and Th contents near these detection limits and must be assessed accordingly.

The limits are imposed principally by low count rates and high backgrounds; increased counting time made little improvement. An additional difficulty which received some attention is background variability. Backgrounds were measured weekly using 330 g NaCl containing 20 ppm K (flame photometer analysis, 1964), and each week's analyses calculated using the average of the backgrounds measured before and after them. When measured daily in July, 1971, the calculated variability was equivalent to .004% K, .036 ppm U, and .081 ppm Th. The variability of weekly averages over several months in 1970 is approximately 25-50% less. Most of this uncertainty is attributable to counting statistics alone, the remaining 10-30% presumably being due to daily variations of natural gamma radiation which may be related to barometric pressure changes. An attempt to monitor these changes by simultaneous background measurements using a 3x3" crystal in the same room and under the same conditions failed, probably due to the much lower counting efficiency of that crystal. (There was greater variation in the factor relating simultaneous background determinations than in the 5x4" crystal background measurements themselves.)

Coefficients of variation, based on the counting statistics for each individual analysis, were less than 15% for $\gtrsim$0.15 ppm U and $\gtrsim$0.5 ppm Th. Reproducibility for a sample (69-816) near these concentration levels is given in Table A4. All samples giving coefficients of variation greater than 15% for any element were re-analyzed and the agreement was usually within 0.1 ppm for concentrations greater than 0.1 ppm.

Accuracy at these low U and Th levels was not established as no appropriate standards were available and neither XRF nor MS7 measurements are appreciably better. I therefore report my results only to the nearest 0.1 ppm when internal coefficients of variation exceeded 15%, or as undetected when below the theoretical detection limits even though there was relative reproducibility below these levels. Numbers should be correct to within 25%.

C. Emission Spectrography

Concentrations of B, Co, Cr, Cu, Ni, Sc, V, and Ba were determined using the techniques of Ahrens and Taylor (1961) and a Jarrel-Ash Ebert Grating Spectrograph Model 71-100. The Pd 3460.8 line was used as an internal standard, the samples being mixed with three times their weight of a C/Pd mixture containing approximately .06% $(NH_3)_4Pd(NO_3)_2$.

Sample weights of 0.1 g were used and analyzed in triplicate. Ilford N.30 photographic plates recorded exposures resulting from arcing electrodes for 2.5 minutes or to completion, and cutting the beam with an eight-step sector having a 2:1 ratio.

The following analytical lines were employed: (in Angstroms) B 2498, Cu 3274, Ni 3414, Co 3453, Sc 4246, Cr 4254, V 4379, and Ba 4554 and 4934. Backgrounds were subtracted where necessary to yield straight line working curves. These were measured by reading plate densities on the low wavelength side of Ni 3414, Co 3453, Sc 4246, and Cr 4254; on the high wavelength side of V 4379; and on the second Si band on the high wavelength side of B 2498. Concentrations were measured by relating line intensities on the photographic plates to beam intensities by use of the Seidel function (Ahrens and Taylor, 1961, p. 142) and graphically relating the ratio of intensity unknown/intensity Pd 3461 to "known" concentrations of standard rocks prepared and arced in an identical fasion. Values adopted for these standards which gave smooth working curves are listed in Table A4.

The precision of this technique for these elements at these concentrations is usually about 5% (e.g. Taylor et al., 1969). Precision measured by 5-9 replicate analyses on separate plates of two Fijian samples is given in Table A4. Coefficients of variation range from 3 to 9% and average 5.9%. Accuracy depends on the working curves but is hopefully comparable to precision. Sr, Ba, Y, and Zr lines were also read for each sample and results compared to values obtained by MS7 and XRF analyses. Agreement was within 10% except for Sr and occasionally erratic Ba. Apparent Ba concentrations above 800 ppm are probably underestimates due to self-absorption.

Lower detection limits taken from Ahrens and Taylor (1961, p. 75) are given in Table A3. Some Fijian samples having apparent concentrations below these levels were reproducible to within 1 ppm or less, and I have reported these concentrations even though comparable standards were not available for verifying their accuracy.

D. Spark Source Mass Spectrography

Concentrations of thirteen rare earth elements (REE), U, Th, Pb, Hf, Cs, Sb, Sn, Nb, Zr, and Y were measured using an A.E.I. MS7 mass spectrograph and the procedures of Taylor (1965, 1971). The latter reference describes in detail the technique I followed. (Analyses appearing in Gill [1970] and Jakes and Gill [1970] did not use the

photoplate processing procedure of Taylor [1971]. Those also used a Lu_2O_3 + C mix prepared and calibrated by myself, using the W-1 and G-1 values cited in those references. All other analyses used Lu Mix E.)

Two to four plates were exposed for each sample involving exposure ranges of 1-1000 x 10^{-9} Coulombs each. Precision, as measured by reproducibility between twelve plates of BHVO-1 basalt arced by A. Graham in 1971 but not processed according to Taylor (1971), is given in Table A4. Individual isotope/Lu176 relative intensities varied from 3 to 10% for most elements in my work, but up to 15% for U, Th, and Pb. Representative values are included in Table A4. Working curves were constructed using values listed in Table A4 for W-1, G-1, AGV-1, and adopted for Apollo 11, 12, and 14 fines. U, Th, and Ba concentrations usually agreed with those obtained by other methods to within 10%. MS7 Ba values were preferred to emission spectrographic results: gamma-ray U and Th were preferred to MS7 data. Cs and Tm were rarely resolvable from C interferences and therefore rarely reported.

A1.5 Sr isotope analyses

Sample aliquots of 100-500 mgm were taken so as to contain 10-20 µgm Sr. These were dissolved and converted to chlorides using 10 ml HF, four drops $HClO_4$, <5 ml H_2O, and <10 ml 2.5 N HCl. Sr was concentrated by passes through two anion exchange columns containing Dowex-50W, X8, 200-400 mesh resin, the first using 20 gm resin and the second, 2 gm. Total Sr blanks per dissolution are typically about 0.02 µgm which would cause error in 87Sr/86Sr ratio determination that is an order of magnitude less than instrumental error.

Isotope ratio measurements followed methods outlined by Compston et al. (1970) and Arriens and Compston (1968). Ratios were calculated using the MASSPEC1 program of P. Arriens.

A modified Nuclide 12" mass spectrometer equipped with a triple filament source, Faraday cup collector, automatic magnetic field peak switching, and digital output was used. Almost all analyses were obtained under the following conditions: 1 to 3 volt stable 88Sr beam intensity; ≦0.2% beam curvature; <0.1% 87Rb; and <0.03% 87Sr tail. To avoid polarization effects a 2 x 10^{10} ohm input resistor was always used when measuring beam intensities. For every sample two or more cycles of 88/86-87/86-88/86 ratio sets were obtained, each set having 15-20 ratios with a coefficient of variation of <0.04% (usually <0.02%) for 87/86

ratios. Ratios were normalized to 88/86 = 8.3752; measured 88/86 ratios ranged from 8.29 to 8.38 and were usually 8.31.

The mean coefficient of variation for populations of 87/86 ratios in 55 individual runs was 0.022 ±.008%, and the mean difference between sets of the same run was 0.00007. The mean difference encountered between nine duplicate runs was also 0.00007, and six runs of the same sample gave ratios differing by 0.00005. The overall precision of 87/86 ratios is therefore thought to be 0.01% or $\pm 7 \times 10^{-5}$ in the samples studied.

Lunar analyses and world-wide studies of 87Sr/86Sr ratio variations have revealed inter-laboratory differences in ratio measurements despite precautions such as those listed above. The E&A shelf $SrCO_3$ was analyzed six times during this study; its 87Sr/86Sr ratio was measured as 0.70817±.00005. Arriens and Compston (1968) found it to average 0.70813 on this machine over a 26 month period. All ratios reported in this thesis have therefore been reduced by 0.00013 to be normalized to 0.7080. The low levels of beam curvature, tail effects, and Rb together rarely exceeded 0.1% so that even were errors made in correcting for them, the effects would be negligible.

These inter-laboratory differences appear due to non-linear electrometer response. The electrometer plus digital measurement system on the mass-spectrometer used in this study has recently been calibrated for linearity using a ramp generator. A ratio of 8.3:1 would be measured too low by about 0.055% so that a bias upwards of 0.0002 in 0.7000 will occur after normalization.

Appendix Two. Geographic and petrographic descriptions of samples studied.

Sample numbers refer to their identification numbers in the ANU rock collection; the year prefices (69- and 71-) are dropped in thesis tables and text. Donated samples are acknowledged; all others were collected personally. Place names were taken from FGSD 1:50,000 scale maps, local informants, and the 1969 Fijian Lands, Mines, and Surveys Department Sheet 1.43 map of Vanua Balavu (Lau).

Possibly ambiguous terms are defined as follows: flow = in situ lava; stream cobble = fresh but not in situ samples from a stream bed; float = fresh but not in situ sample found in the bush. Samples are assigned to the stratigraphic unit appropriate to their locale according to available geological maps, apart from exceptions discussed in the text. More explicit sample locations are available from the author or, for Viti Levu samples, from P. Rodda, FGSD.

The following mineral abbreviations are used: Ab = albite; And = Andesine; Anl = Analcite; Ap = apatite; Biot = Biotite; Byt = bytownite; Calc = calcite; Chab = chabazite; Chal = chalcedony; Chl = chlorite; Cpx = clinopyroxene; DGl = divitrified glass; Ep = epidote; Fs = undifferentiated feldspar; Gl = glass; Hb = hornblende; Kfs = potassium feldspar; Lab = labradorite; Law = lawsonite; Mont = montmorillonite; Ol = olivine; Olig = oligoclase; Opq = undifferentiated opaque; Opx = orthopyroxene; Phil = phillipsite; Pl = plagioclase; Preh = prehnite; Px = undifferentiated pyroxene; Qtz = quartz; San = sanidine; Ser = sericite; Smec = smectite (= hard to identify clay minerals); Sph = sphene; Thom = thomsonite; Zeol = undifferentiated zeolite; and Zir = zircon. K.A.W. Crook helped me identify some zeolites and phyllosilicates. Modes are visual estimates.

Samples 48 to 76 are described elsewhere (Gill, 1970).

69-795 Vanua Levu; Mbua Volcanics(?). E 179°23.5', S 16°24.3'; core from proposed quarry site, Qawa. Porphyritic. Phenocrysts: Pl (20%); fresh Ol (5%); Cpx (7%). Matrix (68%): Pl, Cpx, Opq.

69-797 Vanua Levu; Natewa Volcanics. E 179°17.6', S 16°34.3'; Delaikoro Road. Clast from volcanic conglomerate. Porphyritic. Phenocrysts: Pl (40%); Cpx, about ½ altered (20%). Matrix (35%): Pl, Px, Opq, DGl.

69-798 Vanua Levu; Natewa Volcanics. E 179°19.0', S 16°34.9'; Delaikoro Peak. Center of pillow lava. Vesicular porphyritic. Phenocrysts: Pl (20%). Matrix (80%): Pl, Px, Opq, DGl.

69-799 Vanua Levu: Natewa Volcanics. E 179°18.9', S 16°34.7'; Delaikoro Road. Flow. Porphyritic. Phenocrysts: Pl (10%), fresh Cpx (30%); altered Px (15%). Matrix (45%): Pl, Px, Opq.

69-801 Vanua Levu: Natewa Volcanics. E 179°14.7', S 16°27.3'; Dreketi Road, beyond Tabia Village. Clast in volcanic conglomerate. Porphyritic. Phenocrysts: And (15%), with some calcite replacement; Cpx (2%); Opq (<1%). Matrix (72%): Pl, DGl.

69-804 Vanua Levu: Naroro Andesite. E 179°27.4', S 16°20.1'; Below Segoq fire watch tower. Appears tuffaceous or allotriomorphic with anhedral pieces of And > Hb > Px set in an aphanitic mesostasis.

69-806 Vanua Levu: Natewa Volcanics. E 179°9.1', S 16°36.3'; Creek behind Saivou Village. Flow. Porphyritic and trachytic. Phenocrysts: Lab (30%); Cpx, often altered (5%). Matrix (65%): Pl (70%); Px (20%); Opq (10%).

69-807 Vanua Levu: Natewa Volcanics. E 178°51.1', S 16°36.8'; Dreketi Road at Naibulu Creek. Flow. Porphyritic. Phenocrysts: And (35%); Cpx, often altered to Chl ± Calc (5%). Matrix (60%): Pl, Px, often altered; Opq.

69-808 Vanua Levu: Mbua Basalt. E 178°45.8', S 16°39.5'; Naselesele Falls. Flow. Porphyritic and trachytic. Phenocrysts: Lab (25%); Ol, often altered (10%). Matrix (65%): Pl, Px, Ol, Opq.

69-810 Vanua Levu: Natewa Volcanics. E 179°27.6', S 16°20.1'; Hills east of Bucaisau River, near Neleba Village. Flow. Porphyritic. Phenocrysts: And (25%); Opx, with Cpx reaction rims (4%); Cpx (3%). Matrix (68%): DGl, Pl.

69-811 Vanua Levu: Natewa Volcanics. E 179°30.7', S 16°24.1'; Mouta Road, east of Gegelemunu P.W.D. Clast in laterite; generally altered. Porphyritic. Phenocrysts: altered, sodic Pl (5%). Matrix (95%): Pl, Cpx, Opq, Chl (15%), Chal.

69-812 Viti Levu: Wainimala Group, Tawavatu Tuff. E 178°4.0', S 18°7.6'. Navua River, south of Namuamua; Tuff. Pieces of sodic Pl, Qtz, Px, Chl, Calc.

69-814 Vanua Levu: Natewa Volcanics, Wairuku Formation. E 179°22.5', S 16°46.1'; Roadcut east of Savusavu. Flow. Porphyritic and vesicular. Phenocrysts: And (30%); Cpx (10%); Ol, with Px rims (3%). Matrix (57%): Pl, Px, Opq, DGl, Zeol.

69-815 Vanua Levu: Natewa Volcanics, Savundrondro Formation. E 179°23.7', S 16°45.4'; Savusavu water supply. Dike. Porphyritic and altered. Phenocrysts: Pl, often sericitized but retaining visible zoning (20%); Cpx (1%); Hb (2%). Matrix (70%): Pl, Opq, DGl, Alteration (9%) including Chl, Calc, Ser. Some large (1-2 cm) phenocrysts observed in hand specimen.

69-816 Vanua Levu: Natewa Volcanics, Savundrondro Formation. Same site as 69-815. Flow. Porphyritic and altered. Phenocrysts: Cpx, sometimes with Hb reaction rims (25%); Pl, turbid and sodic (30%). Matrix (45%): Chl, DGl, Zeol (Thompsonite), high-Fe Ep, Calc, Opq, Qtz.

69-817 Vanua Levu: Natewa Volcanics, Lesiatheva Formation. E 179°18.5', S 16°47.9'; Lesiaceva Point. Flow. Porphyritic. Phenocrysts: Lab (28%); Cpx (2%). Matrix (70%): Pl, Px, Opq, DGl, minor Zeol.

69-818 Vanua Levu: Natewa Volcanics, Wairuku Formation. E 179°24.4', S 16°47.3'; Wairuku Creek; type locality. Pillow center. Hypocrystalline, medium-grained, granular. Cpx, often subophitic (55%); Pl (40%); Ol, usually altered (3%); Opq (2%); Gl (tr.).

69-819 Vanua Levu: Natewa Volcanics, Wairuku Formation. E 179°25.5', S 16°47.5'; Creek east of Wairuku. Pillow center. Aphanitic and very altered. DGl, Pl, Px, Zeol, Qtz.

69-820 Vanua Levu: Natewa Volcanics. E 179°43.7', S 16°42.3'; Buca Road, east of Basagasau Village. Clast from volcanic conglomerate. Porphyritic and trachytic. Phenocrysts: Pl (6%); Opx (3%); Cpx (2%); Opq (1%). Matrix (88%): Pl, Px, Opq, minor Zeol. Occasional Hb phenocrysts observed in hand specimen.

69-821 Vanua Levu: Natewa Volcanics, Ndakunimba Formation. E 179°52.0', S 16°42.8'; Roadcut south of Nawi Village. Flow. Porphyritic. Phenocrysts: Pl (10%); fresh Ol (20%). Matrix (70%): Pl, Px, Ol, Opq, Gl.

69-822 Vanua Levu: Naroro Andesite. E 179°7.6', S 16°31.7'; Benuca plug. (From W. Hindle) Porphyritic. Phenocrysts: And (30%) with some alteration to Calc; Opx (5%); Hb (2%) with reaction and oxidation rims; Opq (1%). Matrix (62%): Pl, Px, minor Ap.

69-823 Vanua Levu: Natewa Volcanics. E 178°53.5', S 16°40.1'; Drill core from near Natua. (From W. Hindle) Flow or sill. Porphyritic with strained phenocrysts. Phenocrysts: And (12%), sometimes sericitized; Cpx (5%). Matrix: Pl, Px, Opq.

69-824 Taveuni. E 179°54.5', S 16°56'; Coast road north of Vuna Pt. Flow. Porphyritic. Phenocrysts: Lab (10%); fresh Ol (8%). Matrix (82%): Pl, Px, Ol, Opq.

69-825 Taveuni. E 179°55.5', S 16°55'; Coast road north of Vuna Pt. Flow. Porphyritic. Phenocrysts: Pl (15%); fresh Ol (3%). Matrix (82%): Pl, Cpx, Ol, Opq.

69-826 Taveuni. E 179°58', S 16°52'; Coast road north of Vuna Pt. Flow. Porphyritic and vesicular. Phenocrysts: Pl (10%); Ol (3%) with minor alteration. Matrix (87%): Pl, Px, Opq.

69-827 Taveuni. W 179°59.5', S 16°48.5'; Near Nayalayala archaeological excavation. Flow. Porphyritic. Phenocrysts: Pl (10%); fresh Ol (8%). Matrix (82%): Pl, Px, Ol, Opq.

69-828 Taveuni. W 179°56.5', S 16°44.5'; Coast road north of Niusawa School. Flow or lava tube. Porphyritic. Phenocrysts: And (10%); Ol (8%) with some alteration. Matrix (82%): Pl, Px, Ol, Opq.

69-829 Taveuni. W 179°55.5', S 16°43.5'; Coast road north of 69-828 site. Flow. Porphyritic and vesicular. Phenocrysts: Lab (15%); Ol (10%) with altered rims. Matrix (75%): Pl, Px, Ol, Opq.

69-830 Taveuni. W 179°50.0', S 16°46.5'; Coast road at Waibula River. Stream cobble. Porphyritic. Phenocrysts: Lab (35%) with more oscillatory zoning than other Taveuni samples (An 64 cores - An 49 rims); Ol (2%). Matrix (63%): And (65%); Px + Ol (30%); Opq (5%).

69-831 Vanua Balavu: Korobasanga Basalt. W 178°56', S 17°14'; Coastal point south of Mualevu Village. Flow. Porphyritic. Phenocrysts: And (10%); Ol (2%). Matrix (88%): Pl (57%); Cpx (35%); Opq (5%); Alteration (3%).

69-832 Vanua Balavu: Lau Volcanics. W 178°58', S 17°15'; Uruone Village water supply. Float. Porphyritic. Phenocrysts: Lab (35%); Opx + Cpx (20%); Opq (2%). Matrix (43%): Pl, Px, Opq, DGl.

69-834 Vanua Balavu: Lau Volcanics. W 178°58', S 17°15'; Road north of Levukana. Stream cobble. Porphyritic. Phenocrysts: Lab (35%); Opx + Cpx (10%). Matrix (55%): Pl, Px, Opq.

69-835 Vanua Balavu: Lau Volcanics. W 178°58', S 17°16'; Road north of Nalele. Stream cobble. Porphyritic. Phenocrysts: Lab - And (15%); Cpx (3%). Matrix (82%): Pl, Opq, Px, DGl.

69-836 Cicia: Lau Volcanics. W 179°19', S 17°47'; Tarakua Village. Float. Porphyritic and trachytic. Phenocrysts: Lab (35%); Cpx (5%); Opq (1%). Matrix (59%): Pl, Px, Opq, DGl.

69-837 Cicia: Lau Volcanics. (Same site as 69-836) Float. Porphyritic. Phenocrysts: Lab (20%); Cpx (5%); Opq (1%). Matrix: Pl, Px, Opq, slightly altered DGl.

69-838 Katafaga: Lau Volcanics. W 178°42', S 17°31'; Float near plantation house. Porphyritic. Phenocrysts: Lab (25%); Opx (7%); Opq (1%). Matrix: Pl, Px, Opq, Gl.

69-840 Batiki (Lomaiviti). E 179°9', S 17°46'. Flow. Porphyritic. From F. Coulson. Phenocrysts: Lab (10%); Cpx, euhedral with inclusions of Ol + Pl (35%); Ol, usually altered (3%). Matrix (52%): xline Pl, Px, Opq, Ol, Zeol.

69-841 Nairai (Lomaiviti). E 179°26', S 17°51'. Plug on Naikovu islet. From F. Coulson. Generally fine-grained mixture of Pl, Cpx, Opq, and Gl. A per cent or so of larger Pl + Cpx xls. Trachytic textures.

69-842 Nairai (Lomaiviti). E 179°25', S 17°46'. Flow on north end of island. From F. Coulson. Porphyritic. Phenocrysts: Cpx (27%) with ol inclusions; Ol (10%) with strong iddingsitic alteration; Pl (30%); Opq (3%). Matrix (30%): xline Pl, Px, Ol, Opq with subophitic textures.

69-843 Nairai (Lomaiviti). E 179°25', S 17°48'. Float with/in supposed crater, central east coast. Porphyritic. Phenocrysts: Lab (15%); Cpx (7%); Hb (2%) with some alteration to Calc; Opq (1%). Matrix (75%): Cpx and Opq interstitial between oriented Pl laths.

69-844 White Rock (Yasawa Islands). E 177°14.1', S 17°22.1'. From P. Rodda. Porphyritic. Phenocrysts: Hb (7%) brown pleochroism, slight reaction rim, sometimes quite altered; Pl (5%). Matrix (88%): rectangular And in a poorly xline mesh.

69-845 Viti Levu: Tholo Plutonics. E 177°52.7', S 18°13.9'; Waidradra Creek. Stream cobble. Granular, coarse-grained. Lab (70%); Cpx (28%); Opq (2%). No greenschist minerals.

69-846 Viti Levu: Tholo Plutonics. E 177°53.6', S 18°13.5'; Nacoroga Creek. Stream cobble. Subophitic, coarse-grained. Fresh Lab (80%); Cpx (18%) about 1/3 altered to Hb, Preh, Calc, and Opq; Opq (2%).

69-847 Viti Levu: Tholo Plutonics. E 177°59.3', S 18°10.8'; Waidrata Creek. Stream cobble. Granular, coarse-grained. Lab, locally sericitized (73%); Cpx with incipient Chl alteration (15%); skeletal Opq (2%); Chl and Ser alteration of intersitices (10%).

69-848 Viti Levu: Tholo Plutonics. E 177°44.5', S 18°10.9'; Vunimoli Creek. Stream cobble. Granular, coarse-grained. Pl (90%); often turbid and altered to Ep + Calc; Hb + Cpx (10%) often altered to Chl + Opq.

69-849 Viti Levu: Tholo Plutonics. E 177°44.5', S 18°10.5'; Vunimoli Creek. Stream cobble. Granular, coarse-grained. Pl (65%); Cpx (35%) often altered to Chl + Opq.

69-850 Viti Levu: Wainimala Group, Lokalevu Keratophyre. E 177°51.1', S 18°9.5'; Stream cobble. Porphyritic. Phenocrysts: Pl, pervasively altered and sericitized but with oscillatory zoning preserved (40%); Cpx, locally altered to Chl + Opq (2%). Matrix (58%): Pl + Px + Opq + Alteration (Chl, Ser, Calc).

69-851 Viti Levu: Wainimala Group, Lokalevu Keratophyre. (Same site as 69-850). Flow. Porphyritic. Phenocrysts: Lab (15%), pervasively altered but preserving oscillatory zoning; generally fresh Cpx (3%) but with local Chl + Opq. Matrix (82%): Pl, Px, Opq, Alteration (Chl, Ser, Calc).

69-852 Viti Levu: Tholo Plutonics. E 177°51.9', 18°10.2'; Volosa Creek. Stream cobble. Granular, coarse-grained. Pl (90%), often anhedral and indurated by Px (10%) which is altered to Chl + Opq.

69-853 Viti Levu: Wainimala Group, Numbuonamboto Volcanic Conglomerate. E 177°51.5', S 18°9.7'; Stream cobble, Volosa Creek Porphyritic and vesicular. Phenocrysts: Lab (30%), many strongly altered to Ab + Lau + other Zeol; but some preserving original twinning and zoning; Ol (2%), completely altered. Matrix (68%): Pl, Px, Opq, DGl including recrystallized feldspar. Vesicles contain Chl, Anl, and Smec.

69-854 Viti Levu: Wainimala Group, Lokalevu Keratophyre. E 177°40.5', S 18°3.7'; Near Kubunataba Peak. Flow. Porphyritic. Phenocrysts: Olig (10%), usually sericitized; Qtz (15%) enveloping Olig; Opq (2%). Matrix (73%): Pl, Qtz, Opq.

69-855 Viti Levu: Wainimala Group. (Same site as 69-854) Dike. Granular, fine-grained. Lab (87%), fractured and turbid but with sericitization only at grain boundaries; Cpx (10%) locally altered to Chl + Opq; Opq (3%), both skeletal and in alteration products at interstices.

69-856 Viti Levu: Wainimala Group. E 177°34.2', S 18°3.9'; Navuwai Peak. Porphyry.

69-857 Viti Levu: Wainimala Group, undifferentiated. E 177°33.7', S 18°00.0'; Roadcut near Savai Peak. Clast from volcanic conglomerate. Porphyritic and vesicular. Phenocrysts: fresh And-Lab (8%); Cpx (2%); Ol (1%). Matrix (89%): Pl, Opq, Gl. Vesicles filled with Zeol (Phil ?).

69-858 Viti Levu: Tholo Plutonics. E 177°50.9', S 17°56.1'; Lato Creek, in situ. Granular, coarse-grained. And (45%), frequently sericitized; Qtz (35%); anhedral Kfs (10%); Hb (2%); Opq (2%); Alteration (Chl, Ser, Zeol).

69-859 Viti Levu: Tholo Plutonics. (Same site as sample 69-858). Stream cobble. Granular, coarse-grained. Fresh Byt (55%); unaltered Ol (20%); Cpx ? + Hb + Opq (25%), as complex assemblage which may be multiple generations of amphibole or amphibole overgrowths on Cpx.

69-860 Viti Levu: Wainimala Group. E 177°33.4', S 17°43.8'; Nasasa Creek. Clast from agglomerate. Porphyritic. Phenocrysts: Lab (25%); oxidized Cpx (2%). Matrix (73%): Pl, Px, Opq, Gl.

69-861 Viti Levu: Koroimavua Andesitic Group. (Same site as sample 69-860). Stream cobble. Porphyritic. Phenocrysts: Cpx (25%); Biot (2%); Opq (2%). Matrix (72%): Pl, Kfs ?, Cpx, Biot, Opq, Zeol (vars. Thom and Anl ?), Ap, Gl.

69-862 Viti Levu: Wainimala Group. E 177°33.3', S 17°44.1'; Near Nasasa Creek. Float. Aphanitic. Oriented Pl (60%); Cpx (35%), often quite altered; Opq (5%).

69-863 Viti Levu: Koroimavua Andesitic Group. (Same site as 68-63). Flow. Porphyritic. Phenocrysts: Pl (2%); Cpx (20%); Biot (1%). Matrix (77%): DGl, Zeol, Ap, Fs, Opq.

69-864 Viti Levu: Mba Basaltic Group. E 177°38.1', S 17°48.5'; Nausori Highlands lumber mill quarry. Flow. Porphyritic. Phenocrysts: Lab (25%) with some sericitic alteration; Cpx (1%), Opq (1%). Matrix (73%): Pl, Cpx, Opq, Gl, and minor Zeol (Anl ?) and Calc.

69-865 Viti Levu: Koroimavua Andesitic Group. (Same site as 68-71). Clast from agglomerate. Porphyritic. Phenocrysts: Cpx (25%), Opq (5%), Ap (1%). Matrix (69%): DGl, locally crystallized to Pl + Cpx + Opq; minor Zeol (Anl ?).

Group

69-866 Viti Levu: Wainimala/ undifferentiated. E 177°31.7', S 17°49.7'; Namosi Creek west of Yavuna Village. Dike. Aphanitic and very altered. Pl (60%) partially recrystallized but twinning and zoning retained, much Ser + Calc; Cpx (20%) usually altered to Chl + Ep + Calc + Opq; Opq (8%); general alteration, including Qtz (12%).

69-867 Viti Levu: Tholo Plutonics. E 177°32.5', S 17°49.8'; In situ, Namosi Creek near Yavuna Village. Granular, coarse-grained. Qtz, Pl, Hb + Cpx + Opq intergrowths.

69-868 Viti Levu: Mendrausuthu Andesitic Group, Namosi Andesite. E 178°15.2', S 18°1.2'; Small creek off Waidina River. Stream cobble. Porphyritic. Phenocrysts: Lab (45%); Cpx (12%); Opq (3%). Matrix (40%): DGl, Pl, Px, Opq, Ap.

69-869 Viti Levu: Mendrausuthu Andesitic Group, Namosi Andesite. E 178°14.9', S 18°1.5'; Near Nasirotu Village. Stream cobble. Porphyritic. Phenocrysts: Lab (40%), fractured and sometimes with Calc; Cpx (15%), also with minor Calc; Opq (5%). Matrix (40%): Pl, Px, Opq, Calc.

69-870 Viti Levu: Mendrausuthu Andesitic Group? E 178°13.0', S 18°1.5'; Wainikora Creek. Porphyry. Granular, medium-grained. Lab (82%); Hb (15%), sometimes altered to Calc, Chl, Sph, or Ser; Opq (3%). Acces.: Ap, Zir.

69-873 Viti Levu: Mendrausuthu Andesitic Group, Namosi Andesite. E 178°7.3', S 18°2.9'; Wainigara River. Clast from volcanic conglomerate. Porphyritic. Phenocrysts: Lab (40%); Cpx (15%); Opq (4%). Matrix (41%): Gl, Pl, Opq.

69-874 Viti Levu: Mendrausuthu Andesitic Group, Namosi Andesite. E 178°7.3', S 18°2.9'; Wainigara River. Clast from volcanic conglomerate. Porphyritic. Lab (50%) having less zoning than other Namosi Andesites; Hb (8%) with sharp, unreacted boundaries; Opq (2%). Matrix (40%): Gl, Pl, Hb, Opq.

69-875 Viti Levu: Mendrausuthu Volcanic Group, Namosi Andesite. E 178°7.3', S 18°3.0'; Wainigara River. Stream cobble. Porphyritic. Lab (20%); Hb (15%) all with reaction rims; Cpx (5%) with inclusions of Pl. Matrix (48%): DGl, Pl, Cpx, Ap.

69-876 Viti Levu: Mendrausuthu Andesitic Group, Namosi Andesite. E 178°6.0', S 18°3.5'; Namosi Gap. Stream cobble. Porphyritic. Phenocrysts: Lab (40%) with minor Ser alteration; Cpx (10%); Opq (3%); Hb pseudomorphs (3%) usually composed of Opq and Chl.

69-878 Viti Levu: Wainimala Group ? E 178°5.5', S 18°3.4'; Wainikoroluva River. Stream cobble. Porphyritic but with continuum of grain size from 0.2-2.0 mm. Lab (77%), often turbid and subhedral; Hb (20%) subophitic; Opq (3%) usually associated with Hb.

69-881 Viti Levu: Koroimavua Andesitic Group. E 177°31.8', S 17°42.8'; Nawainiu Creek. Micromonzonite. Granular, medium-grained. Pl (3%); Kfs (74%); Cpx (10%); Biot (5%); Opq (7%); Ap (1%). Kfs often altered to Calc, Lau, and Ep.

69-882 Viti Levu: Koroimavua Andesitic Group. E 177°26.8', S 17°14.2'; Outcrop behind farmer's house at easternmost edge of Sambeto Range. Flow. Phenocrysts: Lab (10%); Cpx (12%) with inclusions of Ap, Biot, and Opq; Opq (3%); Biot (2%); Ap (tr). Matrix (81%): DGl, Pl, Opq, Ap.

69-884 Viti Levu: Mba Basaltic Group. E 177°28.2', S 17°39.4'; Varaqa Creek. Flow. Porphyritic: Lab (45%); unaltered Ol (5%); Cpx (2%); Opq (2%). Matrix (46%): Pl, Cpx, Opq, Biot.

69-885 Viti Levu: Mba Basaltic Group. E 177°34.7', S 17°30.3'; P.W.D. Quarry, King's Road 13 mi. from Lautoka. Flow. Porphyritic. Phenocrysts: Pl (30%) including large (1.5 mm) subhedral clusters of labradorite and small (0.2 mm) laths of andesine; Hb (10%) all with reaction rims and sometimes mantling Cpx; Cpx (3%); Opq (1%). Matrix (56%): Pl, Cpx, Opq.

69-886 Viti Levu: Koroimavua Andesitic Group. E 177°42', S 17°40'; Upstream from 69-887 site. Stream cobble. Porphyritic. Phenocrysts: Lab (30%); Cpx (15%); Opq (4%). Matrix (51%): DGl, Pl, Cpx, Opq. Some alt. to Calc + Ser.

69-887 Viti Levu: Koroimavua Andesitic Group. E 177°42', S 17°40'; Upstream from Balevuto Village on Wasali Creek. Flow. Porphyritic with equigranular, fully crystallized matrix. Phenocrysts: Pl (complete range in size between phenocrysts and matrix); Cpx (15%); Opq (4%). Some Calc and Zeol.

69-888 Viti Levu: Mba Basaltic Group. E 177°42.5', S 17°30.1; King's Road near Talecake Village. Dike. Porphyritic. Phenocrysts: Lab (5%); Cpx (3%); Opq (1%). Matrix (91%): Pl, Px, Opq, DGl.

69-889 Viti Levu: Mba Basaltic Group, Vatia Andesite. E 177°46.2', S 17°25.7'; Vatia Peninsula. Stream cobble. Porphyritic with continuum of grain size between phenocrysts and matrix. Phenocrysts: Pl (15%); Opq (2%). Matrix (83%): Pl, Cpx, some alteration to Zeol, Opq, Chl.

69-890 Viti Levu: Mba Basaltic Group, Vatia Andesite. E 177°46.0', S 17°25.5'; Roadcut, Vatia Peninsula. Flow. Porphyritic. Phenocrysts: Pl (35%) with some Hb inclusion and alteration to Calc; Hb (25%) quite altered all with strong reaction rims. Matrix (40%): Pl, DGl, Px, Zeol, Opq.

69-891 Viti Levu: Mba Basaltic Group, Vatia Andesite. E 177°46.2', S 17°24.4'; Roadside north of mangrove swamp, Vatia Peninsula. Flow. Porphyritic. Phenocrysts: Lab-And (50%); Cpx (15%); Hb (1%) mostly resorbed. Matrix (34%): Pl, Cpx, Opq, DGl.

69-892 Viti Levu: Mba Basaltic Group, Vatia Andesite. (Same site as 68-891) Float. Porphyritic. Phenocrysts: Pl (35%); Cpx (8%); Hb (1%), mostly resorbed; Opq (4%). Matrix (52%): Pl, Px, Opq.

69-893 Viti Levu: Mba Basaltic Group. E 177°59.2', S 17°36.7'; Roadpost 207, Drau Road. Sill. Porphyritic. Phenocrysts: Ol (25%) with minor alteration; zoned Cpx (35%); Opq (3%). Matrix (37%): Pl, Cpx, Opq, Biot, Zeol.

69-894 Viti Levu: Mba Baslatic Group, Tavua Trachyandesite. E 177°52.3', S 17°31.6'; Basala Creek. Flow, Porphyritic. Phenocrysts: Pl (10%); anhedral Cpx (1%); Biot (5%) usually about 1 mm long; Opq (1%). Matrix (81%): Pl, DGl, Opq, Biot, Px, Kfs?, Ap.

69-895 Viti Levu: Mba Basaltic Group. E 177°55.5', S 17°31.4'; Vaga Creek. Float. Porphyritic. Phenocrysts: zoned Cpx (40%) with inclusions of Pl, Ap, Ol along zone boundaries; Ol (10%) usually altered; Lab (5%); Opq (2%). Matrix (43%): Pl, Px, Opq, Ol?.

69-896 Viti Levu: Mba Basaltic Group, Wainivothe Trachybasalt. E 177°55.7', S 17°29.5'; Near Davotu Village. Flow. Porphyritic. Phenocrysts: Lab (45%); Cpx (25%); Opq (2%); Ol pseudomorphs (2%). Matrix (26%): Pl, Px, DGl, Opq, some Calc.

69-897 Viti Levu: Mba Basaltic Group, Wainivothe Trachy basalt. (Same site as 69-896). Float. Porphyritic. Phenocrysts: Cpx (20%); Pl (1%) mostly altered; Opq (1%); Ap (1%). Matrix (77%): Opq, Pl, Cpx, Smec, Phil.

69-898 Viti Levu: Mba Basaltic Group, Wainivothe Trachybasalt. E 177°55.3', S 17°29.5'; Creek near Navotu Village. Aphyric Flow. Some Calc alteration.

69-899 Viti Levu: Mba Basaltic Group, Nakorotumbu Basalt. E 178°14.2', S 17°24.6'; P.W.D. quarry, Nanokonoko Village. Flow. Porphyritic. Phenocrysts: Cpx (20%); Lab (5%); Ol (2%) about ½ altered; Biot (tr); Opq (1%). Matrix (72%): Pl, Cpx, Opq, Biot.

69-900 Viti Levu: Mba Basaltic Group, Nakorotumbu Basalt. (Same site as 69-899). Float. Porphyritic. Phenocrysts: Cpx (30%); Ol (10%); Lab (15%). Matrix (45%): Pl, Cpx, Opq, some Chl after Cpx.

69-901 Viti Levu: Wainimala Group, Wainimbuka Trachyte. E 178°24.0', S 17°42.1'; Waisa Creek. Flow. Porphyritic. Phenocrysts: And (10%); Cpx (1%), some pseudomorphed; Opq (1%). Matrix (88%): trachytic Pl, DGl, Zeol.

69-902 Viti Levu: Mba Basaltic Group. E 178°23.6', S 17°41.8'; Waisa Creek. Stream cobble. Porphyritic. Phenocrysts: Cpx (45%); Ol (15%) often altered to Smec + Chl; Opq (3%). Matrix (37%): DGl, Pl, Biot, Px, Ap, Opq, Thom.

69-903 Viti Levu: Wainimala Group, Korombalavu Andesite. E 178°23.5', S 17°41.5'; Cliff on Waisomo Creek. Clast from volcanic conglomerate. Porphyritic. Phenocrysts: And (12%); Opx (1%); Opq (1%). Matrix (86%): Gl.

69-904 Viti Levu: Wainimala Group, Korombalavu Andesite. E 178°22.4', S 17°41.3'; Base of waterfall, Wairuku Creek. Flow. Porphyritic. Phenocrysts: Lab (35%); Cpx (7%); Opq (1%). Matrix (57%): Pl, Px, Opq.

69-905 Viti Levu: Wainimala Group, Korombalavu Andesite. E 178°23.2', S 17°43.4'; Creek northeast of Nasautoka village. Stream cobble. Aphyric. Pl (70%) probably sodic (Ab-Olig) with definite secondary Ab growing in mesostatis; Cpx (5%) occasionally associated with resorbed Ol; Opq (5%); Chl + Preh + Calc (20%).

69-906 Viti Levu: Wainimala Group. E 178°24.1', S 17°44.2'; Waitoa Creek? Stream cobble. Porphyritic. Phenocrysts: Pl (10%) mostly altered to Ser + Chl, although original zoning preserved. Matrix (90%): Pl, Cpx, Opq, Chl, Ep, Ser.

69-907 Viti Levu: Wainimala Group? E 178°24.1', S 17°45.6'; King's Road at Wailotua Creek. Clast from volcanic conglomerate. Porphyritic. Phenocrysts: Cpx (30%) with minor alteration to Calc + Opq; Pl (25%) quite altered to Ser + Calc; Opq (5%). Matrix (40%): DGl, Pl, Cpx, Opq, Mont, Calc.

69-908 Viti Levu: Tholo Plutonics. E 178°28.0', S 17°46.9'; In situ, Waimaro River. Porphyritic. Phenocrysts: Pl (25%); Hb (30%), with some alteration to Calc + Opq + Pl + Smec; Opq (3%). Matrix (42%): Pl, Opq, DGl.

69-909 Viti Levu: Tholo Plutonics. E 178°29.9', S 17°43.9'; Wailotu River. Stream cobble. Granular, medium-grained. Pl (57%) turbid and sericitized; Qtz (30%); Opq (4%); Chl (6%); and Ep, Calc, Ser.

69-910 Viti Levu: Tholo Plutonics. (Same site as 69-909). Stream cobble. Granular, medium-grained. And-Lab (55%); Hb (7%) altered to Chl or possibly two and different growths of Hb; Qtz (35%); Opq (3%).

69-911 Viti Levu: Mba Basaltic Group, Tova Andesite. E 178°29.5', S 17°35.8'; Waidrou Creek. Stream cobble. Porphyritic. Phenocrysts: Pl (40%); Cpx (15%); Opq (2%). Matrix: Pl, Px, Opq, Biot?.

69-912 Viti Levu: Mba Baslatic Group, Nakorotumbu Basalt. E 178°31.2', S 17°37.4'; North of Luvunivuaka Village. Flow. Porphyritic. Phenocrysts: Cpx (20%); Ol (5%) about 2/3 altered; Opq (2%). Matrix (73%): Pl, Cpx, Ol (altered), Biot.

69-913 Viti Levu: Savura Volcanic Group, Vango Volcanics. (Same site as 68-60). Aphyric flow. Lab (60%); Cpx (35%) usually altered; Opq (5%).

69-914 Viti Levu: Savura Volcanic Group, Vango Volcanics. E 178°26.7', S 18°5.0'; Vago Creek, near water supply. Composite of clasts from conglomerate. Pl and Cpx phenocrysts common. DGl in matrix with some Calc and Zeol.

69-915 Viti Levu: Savura Volcanic Group, Vango Volcanics. E 178°26.7', S 18°4.9'; Beneath waterfall, Vago Creek. Clast from volcanic conglomerate. Aphyric. Pl (60%); Px (5%) usually altered; Opq (3%); DGl (42%). Some Zeol (Chab ?), Chl, and/or Smec.

69-916 Viti Levu: Mba Basaltic Group, Nakombalevu Basalt. (Same site as 69-914). Vago Creek. Stream cobble. Porphyritic. Phenocrysts: Lab (58%); zoned Cpx (10%); Ol (10%) about ¾ altered to iddingsite; Opq (2%). Matrix (20%): Pl, Cpx, Opq, DGl.

69-917 Viti Levu: Mba Basaltic Group, Nakombalevu Basalt. (Same site as 69-915). Stream cobble. Porphyritic. Phenocrysts: Lab (50%); Cpx (8%); Ol (5%); Opq (2%). Matrix (35%): Pl, Cpx, Opq.

69-918 Viti Levu: Mba Basaltic Group, Nandrau Sill. E 178°4.1', S 17°42.4'; Top, Savu Kuita, Wailoa River. In situ monzonite. Granular. And (20%); San (55%); Cpx (12%) often replaced by Chl, Calc, Zeol, Opq; Biot (6%); alteration (7%) usually Chl, Zeol.

69-919 Viti Levu: Mba Basaltic Group. (Same site as 69-918). Stream cobble. Porphyritic. Phenocrysts: Cpx (35%); Ol (4%); Pl (6%); Opq (4%); Biot (2%). Matrix (49%): Pl, Kfs, Px, Zeol, Leucite?, DGl.

69-920 Viti Levu: Wainimala Group, Korombalevu Andesite. E 178°6.7', S 17°44.6'; Wainavi Creek. Pillow center. Porphyritic. Phenocrysts: Lab (25%). Matrix (75%): Pl, Px, Opq. Possible Ol pseudomorphs.

71-366 Viti Levu: Wainimala Group, Tawavatu Tuff. E 178°10.8', S 18°6.3'; Wainatu Creek. Stream cobble. Porphyritic. Phenocrysts: Lab (30%); Cpx (15%); Opq (2%). Matrix (53%): very fine-grained; Pl, Opq visible.

71-367 Viti Levu: Savura Volcanic Group, Nasinu Basalt. Same site as 68-61. Fresh flow sample from quarry blasting. Holoxline. Lab (80%); Ol (10%), Cpx (8%), and Opq (2%) in interstices. Some alteration of ol.

71-368 Vanua Balavu. Korombasanga Basalt. W 178°55.6', S 17°14.0'; Flow exposed along coast. Porphyritic. Phenocrysts: Lab (45%); Ol (8%) about 1/3 altered to iddingsite; Cpx (5%), Opq. Matrix (42%): DGl, Pl, Px, Opq, Ol + alteration.

71-369 Vanua Balavu: Lau Volcanics. W 178°55.1', S 17°13.3'; Float along beach where small ridge intersects seashore. Porphyritic. Phenocrysts: Lab (20%); Opx (3%), Opq (1%). Matrix: Gl, Pl, Opq, Px.

71-370 Vanua Balavu: Lau Volcanics. W 178°55.2', S 17°12.9'; Outcrop on track south of Mavana cemetery. Clast from conglomerate. Porphyritic. Phenocrysts: Lab (10%), Opq (1%). Matrix (89%): DGl, Pl, Opq. Some Zeol alteration.

71-371 Vanua Balavu: Korombasanga Basalt? W 178°55.7', S 17°12.2'; Float beneath prominent cliff along Butalevu Creek. Porphyritic. Phenocrysts: Pl (15%) with Lab cores and And rims; Hb (7%) usually very reacted with only Opq-defined skeletons remaining; Opq (1%). Matrix (77%): finely xline mixture of Pl, DGl, Opq. Contains inclusions of hornblende gabbro surrounding glass. Some Zeol.

71-372 Vanua Balavu: Korombasanga Basalt. W 178°55.9', S 17°12.9'; Cobble in Sosovaka Creek behind Mavana. Porphyritic. Phenocrysts: Pl (40%); Ol (10%), sometimes with iddingsitic alterations; Cpx (10%). Matrix (40%): DGl, Pl, Cpx, Ol, Opq.

71-373 Vanua Balavu: Lau Volcanics. W 178°57.5', S 17°14.5'; Outcrop on track between Muamua Village and Muamua Creek. Clast from conglomerate. Porphyritic. Phenocrysts: And-Lab (30%); Opx (5%); Opq. Matrix (65%): Gl, Pl, Opq.

71-374 Vanua Balavu: Lau Volcanics. W 178°58.7', S 17°15.3'; Flow from outcrop on track south of Saso, protected by mangroves, near Koronibati Point. Porphyritic. Phenocrysts: Pl (20%); Opx (10%), sometimes mantled by Cpx; Cpx (3%); Opq (1%). Matrix: Gl, Pl, Opq.

71-375 Vanua Balavu: Lau Volcanics. W 178°58.1', S 17°17.6'; Clast from agglomerate from easternmost Yanu Yanu, opposite Lomaloma. Porphyritic. Phenocrysts: Lab (12%); Opx and Cpx (2%); Opq (1%). Matrix (85%): Gl, Pl, Px, Opq.

71-376 Vanua Balavu: Korobasanga Basalt. W 178°58.3', S 17°17.6'; Cobble from southwest corner Yanu Yanu. Coarsely granular. Lab (73%); Ol (20%) with some iddingsite; Opq (7%). Possible Zeol.

71-377 Vanua Balavu: Lau Volcanics. W 178°55.0', S 17°12.8'; Outcrop of flow on coastal point southeast of Mavana. Porphyritic. Phenocrysts: Lab (25%); Cpx and Opx (2%); Opq (2%). Matrix (71%): Gl, Pl, Opq. One hornblende gabbro inclusion where hornblende is apparently reacting to Cpx + Opq.

71-378 Vanua Balavu: Korombasanga Basalt. W 178°56.2', S 17°13.5'; Stream cobble just upstream from Mualevu water supply. Porphyritic. Phenocrysts: Lab (50%); Ol (10%) with minor iddingsite. Matrix (40%): Pl, Ol, Opq, DGl somewhat chloritized.

71-379 Vanua Balavu: Lau Volcanics. W 178°55.8', S 17°13.5'; Clast from agglomerate exposed along coast north of Boitaci. Porphyritic. Phenocrysts: Lab (25%); Opx (3%); Opq (1%). Matrix (71%): Gl, Pl, Opq.

71-380 Cikobia: Lau Volcanics. W 178°47.4', S 17°16.9'; Clast from agglomerate on coast southwest of the village. Porphyritic. Phenocrysts: Lab (40%); Opx (5%); Opq (1%). Matrix (56%): DGl probably hydrated, Pl, Px, Opq.

71-381 Cikobia: Lau Volcanics. W 178°46.8', S 17°16.3'; Clast from agglomerate on north side of island, just east of limestone contact. Porphyritic. Phenocrysts: Pl (20%); Opx (5%); Opq (1%). Matrix (74%): Gl, Pl, Opq.

71-382 Vanua Balavu: Lau Volcanics? W 178°57.6', S 17°13.6'; "Odinite" clast from conglomerate at Monicola Point, Porphyritic and quite weathered. Phenocrysts: Cpx (13%) and Opx (2%). Matrix (85%): DGl, Pl, Opq.

71-383 Cikobia: Lau Volcanics. W 178°46.3', S 17°16.4'; Clast from agglomerate outcrop southeast of limestone contact. Porphyritic. Phenocrysts: Lab (20%) with And rims; Opx and Cpx (5%) where Cpx mantles some Opx; Opq (1%). Matrix (74%): Gl, Pl, Px, Opq.

71-384 Munia: Lau Volcanics. W 178°53.0', S 17°22.3'; Clast from very large float, west coast. Porphyritic. Phenocrysts: Cpx (3%); And-Lab (10%); Opq (2%). Matrix (85%): DGl, Pl, Opq.

71-385 Munia: Lau Volcanics. W 178°52.6', S 17°21.5'; Beach cobble at place where track climbs ridge. Porphyritic. Phenocrysts: Pl (15%); Opx (15%); Opq (2%). Matrix (68%): Gl, Pl, Px, Opq.

71-386 Viti Levu: Wainimala Group, Numbuonamboto Volcanic Conglomerate. Same site as 68-48. Flow. Porphyritic. Lab (15%) with minor Ser and Calc; Cpx (2%) about ½ altered to Calo, Chl, Opq; Opq (1%). Matrix (82%): DGl, Pl, Px, Opq.

71-387 Viti Levu: Wainimala Group. Uncertain location. From P. Rodda (sample number C1277). Generally altered. Fs (6%) in matrix of intergrown Fs, Qtz, and Opq with Ap, Ser, and Chl.

71-388 Munia: Lau Volcanics. W 178°52.0', S 17°21.6'; Clast from conglomerate outcrop along east coast road. Porphyritic. Phenocrysts: Pl (20%) with inclusions of fresh glass; Opx (1%); Opq (1%). Matrix (78%): Gl, Pl, Px, Opq.

71-389 Lomaiviti; Welagi Islet, Matuka. E 179°44', S 19°19'. P. Rodda sample C893. Center of flow pillow. Porphyritic. Phenocrysts: Pl (25%); Cpx (5%); Ol pseudomorphs (2%). Matrix (68%): Pl, Cpx, Opq, DGl.

71-390 Ono, Kadavu Group. W 178°28.0', S 18°53.5'. P. Rodda sample C1710. Float from beach west of Nabouwalu. Porphyritic. Phenocrysts: Pl (35%); Hb (8%), quite altered and with strong reaction rims; Cpx (4%), euhedral; Opx (1%); Opq (1%). Matrix (51%); poorly xline mixture of Pl and Cpx.

71-391 Tavua, Mamanuca Islands. W 177°5.5', S 17°36.7'. P. Rodda sample C1378. Dike north of village. Porphyritic. Phenocrysts: And (15%) with some Ser; Hb (8%) with clean margins; Opx (1%); Opq (1%); Ap. Matrix (75%): DGl, Pl, Px, Opq. Also a hornblende gabbro inclusion with very turbid Pl.

71-392 Yavuriba, Mamanuca Islands. W 177°6.2', S 17°30.5'. P. Rodda sample C1398. Clast from conglomerate. Porphyritic. Phenocrysts: Lab (20%); Opx (8%) sometimes mantled by Cpx; Opq (2%). Matrix (70%): Gl, Pl, Opx, Opq. Also two noritic inclusions containing Zeol.

71-393 Ono-i-Lau. W 178°41', S 20°41'. R. Richmond sample P106. Dike beside radio station, Ono Levu. Porphyritic. Phenocrysts: Lab (15%). Matrix (85%): DGl with microlites of And, specks of Px, and Opq.

71-394 Moala, Lomaiviti. E 179°55.0', S 18°35.5'. P. Rodda sample C880. East of Keteira. Porphyritic. Phenocrysts: zoned Cpx (12%); Ol (8%) usually altered along fractures. Matrix (80%): Pl, Cpx, Opq.

B 535 E 179°55', S 16°13'; west of Wainika. Sample from M. Rickard (see Rickard, 1966, p.60). No thin section available.

M 96 Vanua Levu. Udu Volcanics. Uncertain location on Udu Peninsula. Sample from M. Rickard. No thin section available.

Appendix 3. Published Articles

1. Gill (1970). Geochemistry of Viti Levu, Fiji, and its evolution as an island arc. Contr. Mineral. and Petrol. 27, 179-203.

2. Jakes, P. and Gill (1970). Rare earth elements and the island arc tholeiitic series. Earth Planet. Sci. Letters 9, 17-28.

3. Gill and Gorton, M. (1972). A proposed geological and geochemical history of Eastern Melanesia. In: The Western Pacific: Island Arcs, Marginal Seas, and Geochemistry (ed. P. Coleman). W. Austral. Univ. Press. (in press)

4. Gill and Compston, W. (1972). Strontium isotopes in island arc volcanic rocks. In: The Western Pacific: Island Arcs, Marginal Seas, and Geochemistry (ed. P. Coleman). W. Austral. Univ. Press. (in press)

A PROPOSED GEOLOGICAL AND GEOCHEMICAL HISTORY OF EASTERN MELANESIA

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ABSTRACT

There are spatial variations in the geochemistry of island arc volcanism which can be used to infer the subduction polarity of synchronous volcanic rocks. By summarizing and using these variations we demonstrate that the Fiji and Lau Islands faced subduction from the northeast during the Late Miocene. This polarity is also inferred for the New Hebrides during the Early Miocene. We propose the continuity of an ancestral New Hebrides-Fiji-Tonga Ridge which developed during the Eocene between a marginal sea (South Fiji Basin) and the Australia-Pacific plate boundary. The ridge remained continuous until the Early Miocene. Fiji and Tonga have separated from one another during the last 5.5×10^6 years whereas the Fiji-New Hebrides separation may be older. Fragmentation resulted in a transition to almost entirely basaltic or shoshonitic volcanism in Fiji in which K_2O and K/Na seem to have decreased as rifting proceeded. The New Hebrides Ridge itself is currently being split in two in a similar fashion and with similar petrological results.

1. INTRODUCTION

The eastern boundary of the modern Australian plate is a discontinuous series of trenches and volcanic ridges on which the New Hebrides, Fiji, Lau, Tonga, and Kermadec islands lie. This paper proposes a tentative history for that area during the Tertiary, based largely on its exposed geology and on the variation in chemical composition of its volcanic rocks.

Throughout the twentieth century it has been supposed that these ridges were in some way a Tertiary continuation of the progressive eastward accretionary growth of the Australian continent (Suess 1904; Glaessner 1950; Coleman 1967). Each era of geological thought has sought to explain this migration of orogenic focus in terms of the reigning paradigms of its day, and we shall do the same by arguing in terms of plate tectonics, the migration of trenches, and the development of inter-arc basins. By "subduction" we shall mean the thrusting of one plate beneath another; by "subduction zone" we shall mean the interface or fault surface between such plates, including its expression at the Earth's surface in submarine trenches; and by "polarity" we shall mean the direction of dip of this fault and therefore the assymmetric spatial distribution of associated features.

Our attention has focused largely on Fiji because it occupies the central portion of the area and comprises its largest land surface, yet has traditionally been considered its most enigmatic portion. It has also been mapped and studied quite thoroughly by the Fiji Geological Survey Department.

Fiji today is not an active island arc although its lithologies, facies relationships, geochemistry, and geographic position all suggest that it was one throughout much of its history (Dickinson 1967; Rodda 1967; Gill 1970). The art of inferring the geological history of Fiji is therefore an exercise in paleo-plate tectonics and must rely on the principal tools of that trade, such as recognition of: linear volcanic ridges and sedimentary basins; Franciscan-like melanges; ophiolites and alpine ultramafics; glaucophane schists and paired metamorphic belts; and spatial variations in the composition of volcanic rocks. Of these, only the first and last are known in Fiji although the third may be present in Tonga (Ewart et al. in press) and the New Hebrides (Mitchell and Warden 1971). Thus our reconstruction is based mainly on the spatial and temporal variations in geochemistry of volcanic rocks exposed on bathymetrically defined linear ridges.

11. GEOCHEMICAL VARIATIONS IN ISLAND ARCS AND PALEO-PLATE TECTONICS

Most modern island arcs are characterized by large quantities of intermediate and acidic (>53% SiO_2) volcanics as well as basalts. The converse is also true: voluminous intermediate and acidic volcanism is usually associated with a subduction environment. We shall, therefore, regard this situation as "typical" of island arcs without speculating on its genetic reason.

Although most modern island arc volcanoes are concentrated in narrow and remarkably straight bands parallel to their associated trenches, there are many places where some volcanoes occur behind this main volcanic front (eg. SW and NE Honshu, Japan; eastern Aleutians; northern Scotia islands; New Britain; Indonesia;

northern Izu islands). The number of volcanoes and the amount of material extruded decreases with distance from the trench (cf. Sugimura 1968, Fig. 10). The chemical characteristics of volcanism likewise vary with distance from the trench (Kuno 1959, 1966; Sugimura 1968; Jakes and White 1970; Jakes and Gill 1970). A qualitative summary of observed trends taken from the above sources and from our own unpublished data is given in Figure 1. (This model will be developed further elsewhere and an attempt made to quantify it). These variations are thought to be independent of the age and thickness of crust upon which volcanism occurs. It is not our purpose here to assign a genetic significance to these variations but simply to assert them as a description of the present actualistic situation, and thereby to provide a more complete geochemical approach to paleo-subduction than is possible with only the K-h approach of Dickinson (1970, pp. 829-836). Care should be exercised when applying this model. Only suites of rocks that are close enough in age to have been produced during the same subduction configuration and for which there is no evidence of structural dislocation or significant chemical alteration, should be used.

In addition to these spatial zonations, many authors have described analogous temporal variations in the geochemistry of island arc volcanism in which erosional unconformities often separate volcanics of differing compositional affinities (Baker 1968; Jakes and White 1970; Gill 1970; Donnelly et al. 1971). These temporal variations, in which successively younger rocks would appear to have been extruded further from a trench axis by reference to Figure 1, may be due to an

ocean-ward migration of subduction sites and volcanic fronts with time. As this migration occurs, an area of initial tholeiitic volcanism can become further and further removed from the trench, and the character of successive volcanic products at that point can change in the sense illustrated in Figure 1. Failure to recognize this led Gill (1970), for example, to false conclusions about the tectonic evolution of Fiji.

Because Fiji is the central and most enigmatic portion of this area, and because only the products of its volcanism remain as tectonically significant evidence, it is important to stress that those volcanics are not featureless piles of indistinguishable andesite. Instead they exhibit regular geochemical variations some of which are precisely those found in modern, normal island arcs.

If island arc volcanism is characterised by certain spatial variations in magma composition, then these variations may find tectonic usefulness. When one finds, as in Fiji, roughly contemporaneous volcanic rocks in the geologic record which exhibit these same spatial variations and between which there is no evidence of structural dislocation, the simplest explanation is that they are the result of magmatism associated with a subduction zone having the polarity implied by Figure 1. This is not a unique interpretation, but it is the simplest.

III STATUS OF GEOCHEMICAL STUDIES

Good geochemical control for eastern Melanesia has become available during the last three years. There are now over 250 analyses of igneous rocks from Fiji of which three-fourths also contain data for fifteen to forty trace elements. Some of these analyses have been published by Rodda (1969) and Gill (1970), but

the rest will be published elsewhere and are only summarized here. Over 200 analyses, some of which are of poorer quality, are available for the New Hebrides although very few include well-determined trace element data. We have drawn on data summarized by Mallick (this volume), Colley and Warden (in prep.), and our own unpublished material. Data for Tonga is taken from Richards (1962), Ewart et al. (in press and this volume), Melson et al. (1970), and Bauer (1970); and for the Kermadec Islands from Brothers (1970).

IV FIRST PERIOD VOLCANISM: EOCENE TO EARLY MIOCENE

The oldest fauna known from eastern Melanesia is Late Eocene or Tertiary b in age. (Cretaceous fossils have been dredged from the western wall of the Tonga Trench (J. Hawkins, pers. comm.) but these could have been underplated during subduction.) Late Eocene shallow-water fossils but not volcanics have been recognized both in conglomerate clasts (Coleman 1969) and in dredge hauls (D. Karig, pers. comm.) from the New Hebrides Ridge. Identical fauna and lithology is found in the basement of Viti Levu, Fiji (Cole 1960) and Eua, Tonga (Cole 1970). Volcanics are associated with the latter two sites and their composition has been described by Gill (1970) and Ewart et al. (in press) respectively.

Volcanism apparently continued until the Early Miocene in the New Hebrides and Fiji but not in Tonga. Late Oligocene and Early Miocene or Tertiary e to f_{1-2} fauna lie interbedded with volcaniclastics and rare flows on the western islands of the New Hebrides Ridge; the composition of these volcanics has been described by Mitchell (1966), Robinson (1969) and Mallick (this volume). Volcanics of this age are also known from Viti Levu, Fiji (Rodda 1967). The absence of identified Oligocene fauna makes it difficult to know whether volcanism was continuous from Eocene to Early Miocene

or sporadic. There were no apparent changes in magma composition during this period in Fiji (Gill, 1970 and unpublished).

Some aspects of the composition of these first period volcanics are summarized in Table 1. The Eua, Tonga material is least fractionated from chondritic abundances of many elements and Ewart argues that they form the uppermost portion of an alpine ultramafic assemblage. It is important to note, as Ewart does, that such an interpretation is in conflict with most recent speculation about alpine ultramafics in that Ewart does not consider those of Tonga to be oceanic crust formed at a mid-ocean ridge and tectonically emplaced during subduction. The former interpretation would be in conflict with their youthfulness and the low Ti and Zr and higher Sr 87/86 ratios of the Eua basalts relative to those of modern mid-ocean tholeiites. Thus the Tongan material could be products of volcanism behind but near a trench and early in the evolution of the Tonga Ridge, in a fashion suggested by Jakes and Gill (1970) for members of the island arc tholeiitic series.

Data for the Eocene-Early Miocene volcanics of Fiji are based on about twice the number of samples reported by Gill (1970) but differ little from his general summary. These volcanics are more enriched in most large cations than those from Eua. It is not known, however, whether a less enriched assemblage lies unexposed beneath Vanua Levu, Fiji, which has a crustal thickness equivalent to that of Viti Levu (J. Worthington, pers. comm.) or whether a more enriched assemblage occupies a similar position beneath Tonga Tapu or the Ha'apai Group, Tonga.

If Mitchell and Warden (1971) are correct in suggesting contemporaneity between Early Miocene volcanism of the western New Hebrides and the ultrabasics and metamorphic basement of the eastern islands, then a polarity can be inferred in which the sub-

duction zone would lie east of the the New Hebrides Ridge. The composition of these Early Miocene volcanics is erratic but generally richer in alkalis than those of Fiji or Tonga.

Thus, broadly contemporaneous island arc volcanism took place in Fiji, Tonga, and the New Hebrides during the Eocene to Early Miocene. It seems likely that these areas formed a continuous volcanic ridge on the Australian side of a west-dipping subduction zone which marked the boundary between the Pacific and Australian plates at this time. This configuration commenced at about the time when Australia separated from Antarctica (LePichon and Heirtzler, 1968) and the time of westward overthrusting of possible oceanic or inter-arc basin crust onto a once continuous New Caledonia-New Guinea Ridge (Davies and Smith 1970; Lillie and Brothers 1970; Avias, this volume). We therefore suggest that the Australia-Pacific plate boundary and its associated west-dipping subduction complex migrated towards the Pacific during the Early Eocene resulting in the development of an ancestral New Hebrides-Fiji-Tonga Ridge. The apparently synchronous overthrusting along the New Caledonia-New Guinea Ridge suggests that this trench migration accompanied the opening of an inter-arc basin which is now the South Fiji Basin. For a time, either a situation like that of the modern Philippines Sea with two active subduction zones facing the Pacific, or one with westward subduction beneath the New Hebrides-Fiji-Tonga Ridge and westward obduction onto the New Caledonia-New Guinea Ridge, may have existed in eastern Melanesia. There is, however, no evidence that part of New Caledonia migrated eastward together with the trench to form a basement for the new island arc system, as is the presumed custom during evolution of inter-arc basins (Karig, 1971).

V SECOND PERIOD VOLCANISM: LATE MIOCENE

Faunas of Late Miocene or Tertiary f_3 to g or N.17-18 age are associated with basement volcanic rocks on Vanua Levu, Fiji (Rickard 1966; Blow 1967). The composition of these volcanics is described by Hindle and Gill (in prep). Similar faunas are found within marls associated with the Namosi andesites of Viti Levu (Rodda 1967; Blow 1967) the compositions of which have been delineated by Taylor et al. (1969) and Gill (1970). The rocks of unmapped and undated Kandavu may also be of Late Miocene age (Phillips, 1965). Limited analytical data for them is available from Rodda (1969) and Gill (unpublished).

The composition of these volcanics is summarised in Table 2 a-e. This provides an effective north-south geochemical section across roughly synchronous volcanic rocks of Fiji. (Section A-A' on Figure 6). Figures 2,3, and 4 further illustrate geochemical differences between these areas. Despite considerable diversity of Vanua Levu rocks, a marked trend is nevertheless apparent and, when compared with the model given in Figure 1, strongly suggests that these islands faced a subduction zone to their north during the Late Miocene. The concentrations of K, Rb, Ba, Cs, P, Pb, Th, U, and La and the ratios Rb/Sr, Th/U, and La/Yb increase north to south whereas the maximum SiO_2 range and total iron and Y contents and the average Na/K and K/Rb ratios increase south to north.

The basement volcanics of the Lau Ridge are known as the Lau Volcanics (Ladd and Hoffmeister 1945). Their age is disputed. The mollusks and larger foraminifera of the overlying Futuna Limestones have been assigned a Middle Miocene or Tertiary f_3 age and correlated with the Suva Marl of Viti Levu (Ladd and Hoffmeister 1945; H. Ladd and C. Adams, pers. communications, 1971).

The latter formation also contains planktonic foraminifera described by Blow (1967) as N.17-18 or Late Miocene and is generally considered to be that young (Rodda 1967). K-Ar measurements by Gill and McDougall for five samples of the Lau Volcanics from Vanua Balavu, Katafaga, and Cicia all indicate ages within the range 7.5-9.0 x 10^6 years or Late Miocene. (These are part of a larger age determination project the results of which will be published later). If these radiometric ages and the Futuna Limestone-Suva Marl correlation are correct, the older fossils within the Futuna Limestone must have been reworked. Thus the volcanics which form the visible basement of the Lau Ridge seem to be as young as those which dominate Vanua Levu, southeast Viti Levu, and possibly Kandavu. And, by inference, they would overlie a Middle Miocene or older sub-basement.

A similar though less well defined sense of tectonic polarity can also be inferred from the geochemistry of these rocks. Table 2 f-g summarises their composition. We have grouped Cikobia-i-Lau, Munia, and Katafaga together as "eastern islands" and Vanua Balavu and Cicia as "western islands". An east to west increase in K, Rb, Ba, U, La/Yb and Rb/Sr can be seen and a decrease in Na/K, K/Rb, and maximum Fe and Y, suggesting that the subduction zone lay to the east.

Thus there is strong geochemical evidence that the Late Miocene (7-10 x 10^6 years ago) was a period of active island arc volcanism in Fiji and that its associated subduction zone lay to the north and east of the Lau Ridge. Any suggestion that Fiji evolved suigenerically during this period or in isolation from traditional island arc environments having the polarity described (Malahoff, preprint; Chase, in press) must be rejected or modified

as they cannot account for these spatial geochemical variations.

Similar arguments cannot be advanced for the New Hebrides or Tonga. Volcanics associated with a Late Miocene fauna are known only from the eastern New Hebrides islands. The only possible Miocene volcanism recorded in Tonga is the "Younger Volcanic Series" of Eua (Hoffmeister 1932; Ewart et al., in press), a weathered tuffaceous horizon overlying Eocene limestone. Its mineralogy differs from that of Late Miocene volcanics of the Lau Ridge by including hornblende and having very little hypersthene. Nevertheless its correlation with the Lau Volcanics is the only apparent link between the observable geology of Tonga and Lau.

Figure 5 is a tentative proposal for the configuration of the Pacific-Australian plate boundary during the Late Miocene. A similar diagram could be drawn for the first period of volcanism (Eocene-Early Miocene), although its trench and volcanic front might be slightly closer to Australia. This proposal is consistent with the subduction polarity inferred for the Fijian profiles A-A' and B-B' of Figure 6 and also with that inferred for the New Hebrides in the Early Miocene (Mitchell and Warden 1971). It suggests that the Vitiaz Trench and Cape Johnson Trough may be remnant subduction sites. At least the Fiji-Lau-Tonga portion of the diagram is likely to have been the case during this period. Whether the New Hebrides was then where we suggest it was is less certain, and is discussed in Section 8.

VI THIRD PERIOD VOLCANISM: PLIOCENE TO RECENT

The character of volcanism in the New Hebrides and Fiji changed dramatically during the Pliocene and ceased, especially in Fiji, to be that of a typical island arc.

Volcanism of this period in Fiji is almost exclusively basaltic; geochemical features are summarized in Table 3. The province includes shoshonites in northern Viti Levu, analyses of which are given by Dickinson et al. (1968), Rodda (1969, pp. 9-13), and Gill (1970). It also includes quartz-, hypersthene-, and nepheline-normative olivine basalts (some of which carry lherzolite nodules) and differentiates from northeast and southeast Viti Levu, the Lomaiviti islands, southwest Vanua Levu, Taveuni, and the northern Lau islands. At least four centers (Bua, Taveuni, Koro, and Korobasaqa) have yielded basalts with the Ti and Zr contents of mid-ocean varieties. The tholeiites of the Lau Basin (Sclater et al. 1971) could be included in this province.

One of the shoshonite centers has been firmly dated at 4.8-5.3 x 10^6 years old (McDougall 1963), the other yielding erratic K-Ar ages between 4.8-9.0x10^6 years (Rodda et al. 1967). Lacking rigorously consistent data for volcanism older than 5 x 10^6 years ago at Tavua or other basaltic centers of this province, we take that figure as the approximate time of initiation of third period volcanic activity. It continued until almost 3x10^6 years ago in the northern Lau islands (Gill and McDougall, unpublished), until about 2x10^3 years ago in Taveuni (E. Frost , pers. comm.), and to as recently as 1948 at Niua'fou in the Lau Basin (Richard 1962). The geomorphology of Viti Levu has been interpreted as indicating over 800 meters of regional uplift since the Pliocene (Dickinson, unpublished manuscript), or within the last 5.5x10^6 years.

Volcanism of this period in the New Hebrides can be divided into two groups. During the early Pliocene high-K rocks were erupted on a ridge extending from Epi south to Efate and possibly Tanna. They include little-known plutonics from the basement of Tanna, high-K calc-alkaline andesites from Western Epi in which

$K_2O/Na_2O \sim 1$, and the trachytes to dacites of Efate with similar alkali ratios (Ash 1970, Colley and Warden, in preparation).

In sharp contrast, basalts and minor andesites characterize the Pleistocene to Recent volcanism, which extends the length of the group in a well defined line lying above 200 km earthquake epicenters. Tanna, which is associated with deeper earthquakes, is exceptional in producing strongly alkali-enriched basaltic andesites. Elsewhere rocks range from tholeiites to alkali basalts and associated andesites.

Because the Pliocene to Recent volcanism of Fiji and the New Hebrides is so unlike what preceded it and what is normal in island arcs, we interpret that volcanism and the simultaneous regional uplift as being related to fragmentation of the New Hebrides-Fiji-Tonga Ridge, rather than simple subduction. This is also a non-unique interpretation as similar low-Ti basaltic volcanism dominates the present-day New Hebrides and New Georgia which are spatially conjoined with subduction. Not until the genesis of island arc magmas is better understood can one be certain of the relationship between subduction and the third period volcanics of Fiji or the New Hebrides.

The overall change in character from oldest shoshonites to youngest low-K tholeiites is similar to the change in volcanic character which apparently accompanies rupturing of continents preceding and during drift (cf. Green 1971, pp. 718-720). Volcanism in Fiji began on the ridge itself and was later concentrated in area C of Figure 6 - that area most likely to be affected by counter-clockwise rotation of the northern Lau Ridge.

If this interpretation is correct, normal subduction volcanism (i.e. that in which intermediate and acidic material figures prominently) ended in Fiji approximately 7x10^6 years ago and disruption of the volcanic ridge formed thereby had begun by 2x10^6

years later. Second, it implies that the locus of basaltic volcanism moved during this period in the direction of the modern trenches to the east and west of Fiji as there is neither historically active volcanism in nor seismic attenuation beneath modern Fiji whereas both appear to be present in and beneath the Lau Basin and Fiji Plateau (Barazangi and Isacks, in press; Sclater et al., in press; Chase, in press). Third, it suggests an explanation for the K_2O variations within the modern New Hebrides (Mallick,this volume) that has no significance for arc polarity. The variations in composition of third period volcanics from Fiji (Table 3) are thought to reflect variations in source material (base of the crust perhaps for shoshonites; heterogeneous upper mantle for basalts) and degree of fusion, rather than depth to an associated Benioff zone.

Third period volcanism of Tonga and Kermadec is dominantly basaltic-andesite to dacite and typical of the island arc tholeiitic series (Ewart et al., in prep.; Bauer 1970; Richard 1962; Brothers 1970). Its tectonic environment is almost archetypal for a normal intra-oceanic island arc.

VIII DATE OF FRAGMENTATION OF THE NEW HEBRIDES-FIJI-TONGA RIDGE

We have proposed a former continuity of the presently fragmented volcanic ridges on the eastern boundary of the Australian plate which persisted through at least the Early Miocene. Our reconstruction (Figure 5) requires a reversal of tectonic polarity and about 45° clockwise rotation of the New Hebrides Ridge, creation of new crust forming the Fiji Plateau, and simultaneous subduction of an ancestral North Fiji Basin beneath the migrating New Hebrides trench and arc. The idea of continuity has been suggested in various forms by several people (Hess and Maxwell

1953; Carey 1963; Coleman 1967; Karig and Mammerickx, in press); the principal remaining question is when the disruption began. There are two options: Middle Miocene or Pliocene to Recent.

An Early Middle Miocene or Tertiary f_3 (9-12.5 x 10^6 years; Page and McDougall 1970) separation of the New Hebrides from Fiji appears more consistent with New Hebridean geology as it is the time of greatest preserved deformation and faulting there. Robinson (1969) reports evidence that strong east-west compressional stress affected Santo and Malekula during this peridd and Rodda (1967) suggests it was also a time of uplift and deformation in Fiji. This may also have been the time when New Guinea as the leading edge of Australia, over-ran a westward extension of the New Britain Trench (Page 1971) and when the Ontong Java Plateau of the Pacific Plate impinged upon the Solomon island ridge (Moberly, this volume).

If instead, separation and polarity reversal commenced in the Pliocene, or less than 5.5 x 10^6 years ago, it would coincide with the change in volcanism noted above for Fiji and the New Hebrides. Moreover, if this event was a passive response to some major re-adjustment of Pacific-Australian plate relationships, it is unlikely that fragmentation would have commenced sooner in the area west of Fiji than east of it. This implies comparable ages for the Fiji Plateau and Lau Basin. The Lau Basin must postdate the age of volcanics forming the basement of the Lau Ridge. If the Late Miocene K-Ar dates mentioned above for these volcanics are correct, the Lau Basin and by inference the Fiji Plateau are no older than Pliocene in age. Such youthfulness for the Fiji Plateau encourages speculation that its anomalous deep earthquake foci, which shoal to the east (Santo 1970), may originate within detached, foundering litho-

sphere dating from the earlier subduction configuration of opposite polarity. Likewise the 250-300 km depths attained by foci near the New Hebrides Trench may represent the depth to which lithosphere has been downthrust so far during the current configuration. If so, lithosphere is discontinuous and deep focus S waves should be attenuated even at seismic stations near the New Hebrides Trench.

The limitation of recognized Late Miocene volcanism to the eastern New Hebrides could indicate either that reversal of subduction polarity had already occurred by then or that minor eastward trench migration had taken place, as was inferred for Fiji.

Karig (1974 and in press) has proposed that the central "Y"-shaped part of the New Hebrides has resulted from rifting apart of the two arms of the "Y". When these two sections are joined by matching bathymetric contours, (Figure 7), the suture thereby defined coincides with a major fault cutting Epi and with the Teuma Graben of Efate. Pliocene volcanics have been uplifted 700-900 m to the west of the former (Warden 1967) and considerable post-Pliocene uplift is also apparent in the latter (Williams and Warden 1964). This was also a time of general uplift in the central New Hebrides.

The volcanoes of Aoba and Ambrym, which lie within this proposed rift, almost exclusively produce rocks which are transitional between alkali and tholeiite basalt. These may be more related to the extensional environment than to normal island arc processes although they have TiO_2 contents typical of island arcs. The K_2O variation described from the Banks Islands (Mallick, this

volume) may reflect the rifting (as we discussed in Section 6) rather than differential depth to the Benioff Zone.

In migrating westward, Santo and Malekula may have overridden the central New Hebrides Trench, thus accounting for its disappearance in this region despite the continuity of volcanism and seismicity. East of Santo earthquakes having focal depths of about 150 km extend for some distance normal to the arc, indicating a flattening of the Benioff Zone.

IX CONCLUSIONS

Detailed knowledge of the geochemistry of island arc volcanic rocks can be a useful tool in paleo-plate tectonic studies, espically in areas such as eastern Melanesia where other criteria by which to establish the site and polarity of former subduction zones are absent. This is strictly true,however, only if a model such as that of Figure 1 represents all modern examples, if uniformitarianism applies, and if the model is independent of crustal thickness and age and grossly variable fractionation processes.

We have distinguished three periods of igneous activity in eastern Melanesia. The first, covering the Eocene to Early Miocene, was predominantly submarine and witnessed the gradual formation the New Hebrides-Fiji-Tonga Ridge east of an early Tertiary marginal sea (South Fiji Basin) and behind the newly established boundary between the Australian and Pacific plates. In Tonga and Fiji the volcanics of this period belong to the island arc tholeiitic series. The second period, separated from the first by an erosional unconformity in Fiji and the New Hebrides and confined to the Late Miocene ($7\text{-}10 \times 10^6$ year ago), is best developed in Fiji. There a polarity can be inferred from the geochemical systematics of volcanic rocks, suggesting southwestward subduction of the Pacific plate.

Volcanism of this period is recorded in the eastern New Hebrides and possibly on Eua, Tonga, but polarity can be inferred for neither.

We suggest that the eastern boundary of the Australian plate involved west-dipping subduction and a normal, continuous island arc during the first and possibly both of these periods. It may have looked something like what is proposed in Figure 5. We attribute the present distribution of islands to fracturing of this island arc during the Middle Miocene and/or Pliocene to Recent, caused or at least accompanied by a different resolution of stresses between Pacific and Australian Plates. This fracturing resulted in shoshonitic and basaltic volcanism in Fiji. The newly established configuration includes normal subduction volcanism of the island arc tholeiitic series in the Tonga and Kermadec chains, and tholeiitic basalt volcanism in the Lau Basin and probably Fiji Plateau. Continued rifting is splitting the New Hebrides Ridge itself, resulting in dominantly basaltic volcanism of diverse kinds.

FIGURE CAPTIONS

Figure 1. Geochemical variations in island arcs. See discussion in text. $^{87}Sr/^{86}Sr$ ratios also decrease with distance from the trench (cf. Gill and Compston, this volume).

Figure 2. Fiji (Late Miocene) K_2O-SiO_2 variations. Data are from Rodda (1967), Hindle and Gill (in prep.) for Vanua Levu, and Gill (in prep.) for Namosi and Kandavu.

Figure 3. Fiji (Late Miocene) Mg-Fe variation curves. Data are summarized from the sources listed beneath Figure 2 and Gill (unpublished) for Lau. Stars indicate the widely differing analyses for two Kandavu samples (the upper point represents a 1918 analysis). Vanua Levu data is more scattered than this curve indicates.

Figure 4. Fiji (Late Miocene) REE patterns. Data are from Taylor et al. (1969), Hindle and Gill (in prep.), and Gill (in prep. and unpublished). Shaded fields represent the range of five Vanua Levu and four Namosi samples, respectively.

Figure 5. A proposed configuration of the New Hebrides-Fiji-Tonga Ridge before fracturing. Ridge outlines represent their modern two km isobaths although the Tonga portion is our estimate of its pre-Pliocene shape, i.e. the modern "non-volcanic ridge". The New Hebrides Ridge should be similarly modified; see Figure 7. The southern Lau Ridge and the Vitiaz Trench have been fixed in their present position for reference in this reconstruction; the Tonga Ridge has been rotated counter-clockwise (although opening of the Lau Basin and Havre Trough cannot be explained by simple rotation about a point in the Taupo region of New Zealand); the Fiji islands rotated clockwise; and the New Hebrides Ridge rotated counter-clockwise about a point amidst the Santa Cruz islands. The patterned area forming a continuation of the Vitiaz Trench is the inferred site of

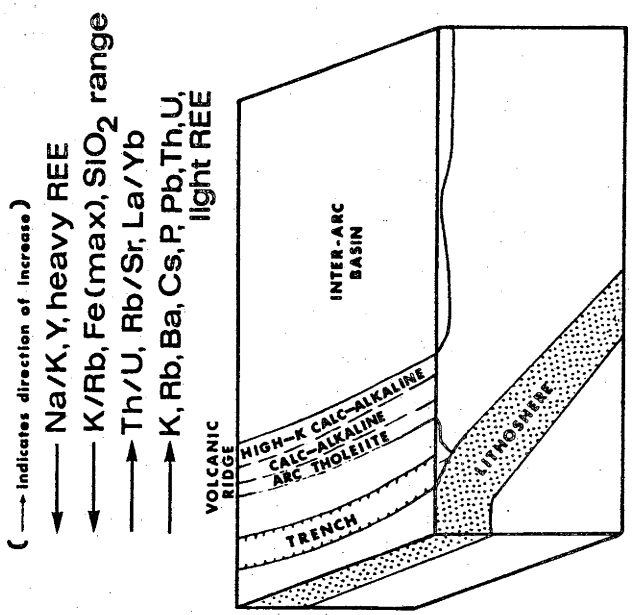
crustal subduction at this time. The 625 km contour of modern earthquake foci is taken from Karig (1971, Figure 4) and is very speculative west of Fiji.

The apparent geometric fit, inferred subduction polarity, youth of ocean basins which are predicted by this model to be <8 x10^6 years old, and fortuitous alignment of the Vitiaz Trench and some modern deep earthquakes, support this reconstruction.

Geographic abbreviations are: M=Maewo; P=Pentecost; ES=Espiritu Santo; Ml=Malekula; Ef=Efate; Vt=Viti Levu; K=Kandavu; Vn=Vanua Levu; L=northern Lau islands; and E=Eua.

Figure 6. Location of cross sections discussed in the text. Area C is meant to include the Lomaiviti islands and eastern Viti Levu.

Figure 7. Left: The present-day New Hebrides including 1500 meter and 5000 meter (shaded) contours. Arrows indicate inferred break-up. Right: Reconstruction of New Hebrides before rifting. The New Hebrides Trench has been restored and the present positions of Santo and Malekula are dotted. Note alignment of the west coasts of Maewo and Pentecost with the central fault on Epi and the Teuma Graben on Efate. A preliminary K-Ar date of about 0.6 x10^6 years has been obtained by one of us (MG) on a fresh but glassy sample from a formation on west Epi which is cut by the central fault and which therefore predates rifting.



2

▫ Vanua Levu (Natewa & Undu)
• Viti Levu (Namosi)
⊙ Kandavu
%K2O
4
3
2
1
50
60
70
80
% SiO2

3

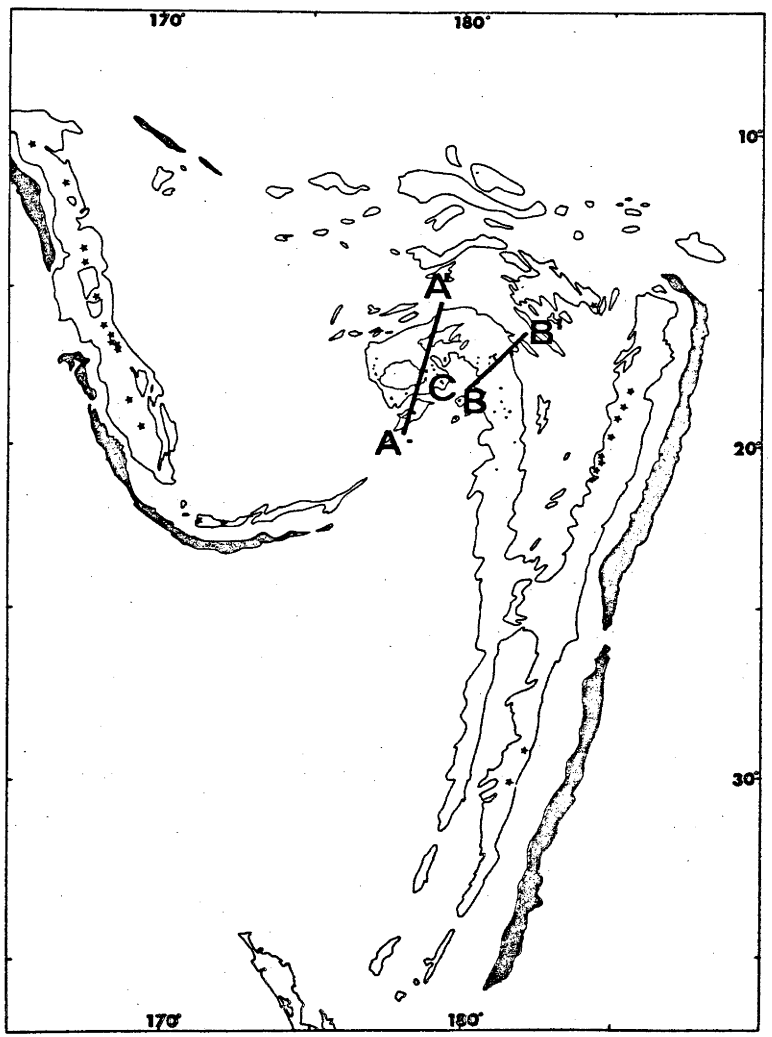
% MgO
8
6
4
2
% FeO + Fe2O3
2
4
6
8
10
12
Vanua Levu
Lau West
Lau East
Namosi

5

Subductive Maiden
recent 625 km
LAU
TONGA
New Hebrides
8-10 m.y.

4

La Ce Pr Nd Sm Eu Gd Tb Dy Ho Er Yb
Rock/Chondrites
10
100
Vanua Levu
Kadavu
Namosi


170°
180°
10°
20°
30°
A
A'
B
B'
C
170°
180°

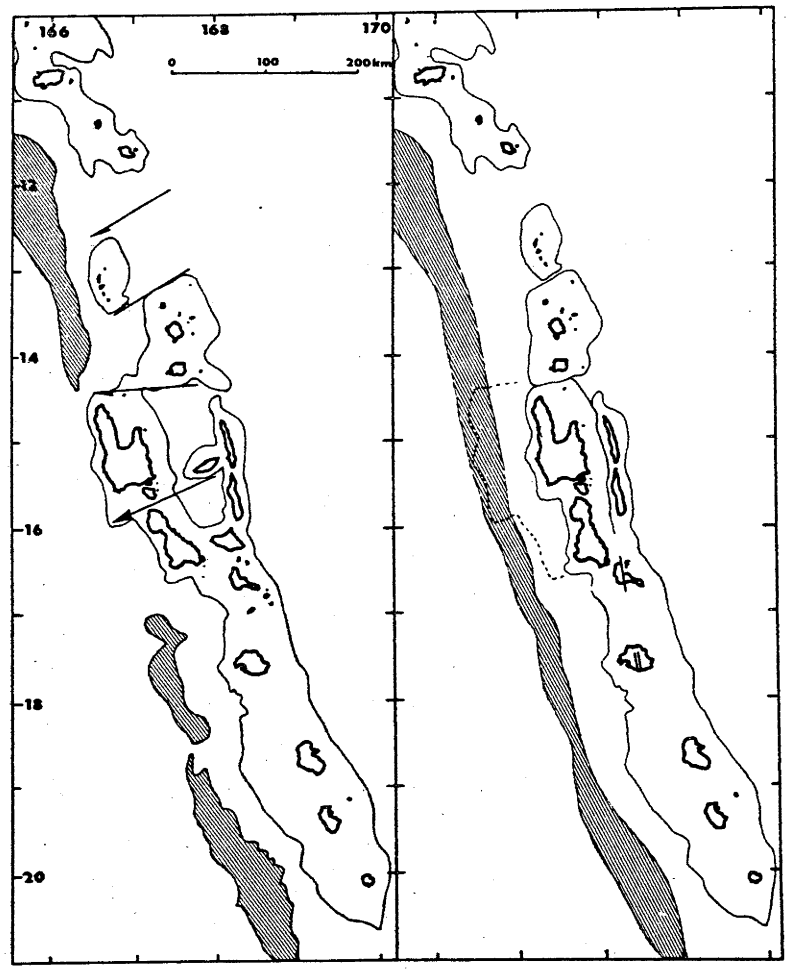

166
168
170
0
100
200km
12
14
16
18
20

TABLE 1

REPRESENTATIVE ANALYSES OF FIRST PERIOD VOLCANIC ROCKS

	Tonga		Fiji			New Hebrides		
	Eua-7*	Eua-11*	69-920	68-60**	68-50	18***	19***	36***
	a	b	c	d	e	f	g	h
SiO_2	50.11	61.35	49.72	59.22	73.31	52.58	62.59	68.26
TiO_2	0.43	0.77	1.07	1.08	0.51	0.91	0.53	0.79
Al_2O_3	21.00	17.41	18.37	15.13	13.13	18.35	16.28	16.01
Fe_2O_3	2.98	3.36	11.59*	8.47	1.59	4.26	4.38	1.59
FeO	5.74	3.46	-	2.54	2.09	5.10	2.26	2.06
MnO	0.18	0.10	0.19	0.15	0.12	0.19	0.10	0.04
MgO	6.04	2.94	4.64	3.50	0.78	3.76	1.40	0.57
CaO	11.39	6.06	11.50	6.17	2.65	9.93	7.07	4.86
Na_2O	1.23	4.09	2.31	3.32	4.78	3.53	3.65	4.15
K_2O	.24	0.32	0.44	0.30	0.91	0.97	1.27	1.51
P_2O_5	.04	0.14	0.16	0.13	0.12	0.30	0.33	0.16
Rb	1.2	2.3	5.1	2.1	6.8			
Sr	115.	145.	280.	130.	116.			
Ba	14.	50.	56.	70.	218.			
Pb	1.	1.	1.	2.	1.			
Th	-	-	0.3	<0.3	<0.3			
U	-	-	0.2	0.1	0.2			
Zr	20.	120.	55.	68.	71.			
Hf	-	-	-	1.5	4.5			
La	0.8	-	-	2.4	2.7			
Yb	1.2	-	-	2.5	6.9			
Y	12.	34.	17.	34.	43.			
Ni	25.	2.	10.	nd	nd			
Co	30.	14.	32.	27.	nd			
Cr	75.	7.	50.	2.	nd			
Sc	34.	18.	34.	35.	16.			
V	230.	120.	266.	190.	9.			
Cu	25.	67.	98.	77.	nd			
Sr87/86	0.7034	0.7039	.7032	.7037	.7048			

* Ewart et al. (in press)
** Gill (1970)
*** Mitchell (1966)

All analyses in this and following tables are re-calculated to anhydrous equivalents. Fe_2O_3*=total iron as Fe_2O_3. Sr 87/86 ratios are normalized to 0.7080 for E and A standard Sr. "nd" means not detected; other omissions mean not determined. Data are from our unpublished studies except where referenced.

TABLE 2

REPRESENTATIVE ANALYSES OF SECOND PERIOD VOLCANIC ROCKS

	FIJI					LAU		NEW HEBRIDES		
	Vanua Levu			Viti Levu	Kandavu	West	East			
	69-818	69-810	M96	X96*	71-390	71-374	71-385	610	626	593
	a	b	c	d	e	f	g	h	i	j
SiO_2	50.04	56.95	79.37	58.60	60.64	55.34	56.64	52.39	52.25	61.69
TiO_2	0.74	0.85	0.20	0.66	0.62	0.96	0.93	0.85	0.76	0.72
Al_2O_3	15.53	16.75	10.42	18.07	17.08	17.33	16.86	16.95	17.54	14.71
Fe_2O_3	11.65*	8.86*	2.49*	3.27	5.92*	8.90*	8.75*	5.26	4.91	3.70
FeO	-	-	-	3.37	-	-	-	4.64	5.01	3.13
MnO	0.19	0.15	0.05	-	0.11	0.17	0.16	0.20	0.17	0.09
MgO	6.73	4.15	0.28	3.52	2.18	4.63	4.24	5.61	4.78	2.80
CaO	12.25	8.49	1.76	7.55	5.74	8.63	8.16	9.46	9.59	6.74
Na_2O	2.12	3.02	4.08	3.47	4.47	2.83	3.27	3.34	3.07	3.25
K_2O	0.59	0.65	1.17	1.28	2.89	1.00	0.79	1.12	1.67	2.85
P_2O_5	0.15	0.13	0.18	0.21	0.34	0.21	0.20	0.19	0.25	0.31
Rb	16.	6.	6.	22.	51.	15.	9.			
Sr	390.	170.	95.	454.	1330.	266.	304.			
Ba	285.	80.	119.	337.	768.	135.	125.			
Pb	2.	1.	3.	2.	8.	2.	2.			
Th	<0.3	0.3	0.4	1.4	4.2	0.6	0.6			
U	0.2	0.2	0.5	.50	1.7	0.6	.45			
Zr	56.	95.	134.	97.	145.	-	64.			
Hf	0.6	1.4	2.0	1.3	2.3	1.5	1.5			
La	3.0	7.0	6.6	11.1	27.5	7.0	9.8			
Yb	1.3	2.4	2.8	1.4	1.4	2.4	4.3			
Y	21.	31.	33.	18.	17.	29.	59.			
Ni	45.	8.	nd	8.	6.	17.	14.			
Co	46.	23.	nd	16.	18.	25.	23.			
Cr	148.	10.	3.	37.	8.	31.	20.			
Sc	46.	27.	9.	18.	12.	27.	36.			
V	300.	208.	nd	163.	166.	214.	220.			
Cu	140.	36.	6.	31.	59.	50.	40.			
Sr87/86	.7036	.7041	.7042	.7036	.7032	-	.7033			

* Taylor et al. (1969)

TABLE 3

REPRESENTATIVE ANALYSES OF THIRD PERIOD VOLCANIC ROCKS

	Tonga-Kermadec			Fiji			New Hebrides			
	10415* a	1** b	39** c	68-64*** d	71-829 e	69-841 f	680 g	516 h	538 i	698 j
SiO_2	49.0	53.68	66.77	51.78	47.80	53.30	61.06	50.73	49.86	51.02
TiO_2	1.2	0.57	0.61	0:63	1.83	0.76	0.87	0.89	0.89	0.71
Al_2O_3	18.3	17.74	14.43	20.27	16.89	19.42	15.25	16.37	16.57	14.00
Fe_2O_3	1.9	2.65	1.08	3.68	9.57*	9.38*	2.23	4.37	4.59	3.62
FeO	6.6	7.56	6.91	3.72	-	-	5.40	6.41	6.45	6.80
MnO	0.2	0.19	0.19	0.15	0.16	0.20	0.16	0.21	0.20	0.19
MgO	7.8	4.25	1.57	3.77	8.86	3.77	2.36	5.92	5.74	8.52
CaO	12.4	11.06	3.95	9.24	10.16	8.90	5.12	10.69	10.93	12.02
Na_2O	1.8	1.80	3.13	2.96	3.33	3.06	3.39	3.02	2.99	2.06
K_2O	0.4	0.40	1.15	3.34	0.95	0.74	3.65	1.16	1.80	0.86
P_2O_5	0.15	0.08	0.19	0.45	0.43	0.47	0.49	0.22	0.34	0.20
Rb		6.	17.	56.	21.	18.				
Sr		235.	300.	1230.	465.	512.				
Ba		100.	190.	677.	275.	360.				
Pb		4.	17.	8.	2.5	3.				
Th		-	-	1.3	2.7	1.7				
U		-	-	0.56	.67	.46				
Zr		28.	49.	46.	146.	124.				
Hf		-	-	1.1	-	-				
La		2.1	5.9	8.5	-	-				
Yb		0.95	1.6	1.5	-	-				
Y		17.	24.	12.	28.	26.				
Ni		13.	nd	13.	170.	11.				
Co		28.	13.	21.	43.	24.				
Cr		15.	4.	40.	402.	5.				
Sc		35.	25.	14.	30.	11.				
V		310.	95.	220.	189.	130.				
Cu		145.	26.	200.	57.	35.				
Sr 87/86		.7038	.7043	.7036	.7035	.7034				

* Brothers and Martin (1970)
** Ewart et al. (in prep.)
*** Gill (1970).

REFERENCES

Ash, R.P. 1970. Annual report of the Geological Survey 1968. New Hebrides Geol. Survey Report.

Avias, J. (this volume). Major features of the New Guinea - Louisiade - New Caledonian-Norfolk Arc system.

Baker, P.E. 1968. Comparative volcanology and petrology of the Atlantic island-arcs. Bull. Volcanol. 32, 189-206.

Barazangi, M. and Isacks, B. (in press). Lateral variations of seismic wave attenuation in the upper mantle above the inclined earthquake zone of the Tonga island arc: deep anomaly in the upper mantle. J. Geophys. Res.

Bauer, G.R. 1970. The geology of Tofua island, Tonga. Pac. Sci. 24, 333-350.

Blow, W.H. 1967. Proc. First Plankt. Conf. Geneva. Vol. 1, p. 305.

Brothers, R.N. 1970. Petrochemical affinities of volcanic rocks from the Tonga-Kermadec island arc, Southwest Pacific. Bull. Volcanol. 34, 308-329.

Brothers, R.N. and Martin, K.R. 1970. The geology of Macauley Island Kermadec Group, south west Pacific. Bull. Volcanol. 34, 330-346.

Carey, S.W. 1963. The assymetry of the Earth. Aust. J. Sci. 25, 369-382 and 479-488.

Chase, C.G. (in press). Tectonic history of the Fiji Plateau. Bull. Geol. Soc. Am.

Cole, W.S. 1960. Upper Eocene and Oligocene larger foraminifera from Viti Levu, Fiji. U.S. Geol.Surv. Prof. Pr. 374-A.

Cole, W.S. 1970. Larger foraminifera of Late Eocene age from Eua, Tonga. U.S. Geol. Surv. Prof. Pr. 640-B.

Coleman, P.J. 1967. A possible resolution of the Melanesian Re-entrant. Upper Mantle Proj.: Second Australian Progress Report. 192-194.

Coleman, P.J. 1969. Derived Eocene larger foraminifera on Maewo, Eastern New Hebrides, and their south-west Pacific implications. 1967 Ann. Report New Hebrides Geol. Surv. 36-37.

Colley, H., and Warden, A.J. (in press). The petrology of the New Hebrides.

Davies, H.L., and Smith, I.E. 1970. Geology of Eastern Papua: a synthesis. B.M.R. Records 1970/116.

Dickinson, W.R. 1967. Tectonic Development of Fiji. Tectonophysics. 4, 543-553.

Dickinson, W.R. 1970. Relations of andesites, granites, and derivative sandstones to arc-trench tectonics. Rev. Geophys. Space Phys. 8, 813-860.

Dickinson, W.R. (unpublished manuscript). Dissected erosion surfaces in Northwest Viti Levu, Fiji. On file with Fiji Geological Survey Dept.

Dickinson, W.R., Rickard, M.J., Coulson, F.I., Smith, J.G., and Lawrence, R.L. 1968. Late Cenozoic Shoshonitic lavas in north-western Viti Levu, Fiji. Nature. 219, 148.

Donnelly, T.W., Rogers, J.J.W., Pushkar, P., and Armstrong, R.L. 1971. Chemical evolution of the igneous rocks of the Eastern West Indies: an investigation of thorium, uranium, and potassium distributions, and lead and strontium isotopic ratios. G.S.A. Mem. 116.

Ewart, A., Bryan, W.B., and Graham, A.L., (in press). The petrology and geochemistry of the igneous rocks from Eua, Tonga. Bull. Geol. Soc. Am.

Gill, J.B. 1970. Geochemistry of Viti Levu, Fiji, and its evolution as an island arc. Contr. Mineral. Petrol. 27, 179-203.

Glaessner, M.F. 1950. Geotectonic position of New Guinea. Bull. Am. Assoc. Petr. Geol. 34, 856-881.

Green, D.H. 1971. Composition of basaltic magmas as indicators of conditions of origin: application to oceanic volcanism. Phil. Trans. Roy. Soc. Lond. A. 268, 707-725.

Hess, H.H. and Maxwell, J.C. 1949. Major structural features of the southeast Pacific: a preliminary interpretation of A.O. 5484, bathymetric chart. New Guinea to New Zealand. Proc. Pacific Sci. Congr. Pacific. Sci. Assoc. 7th, New Zealand, 1949. 2: 14-17.

Hoffmeister, J.E. 1932. Geology of Eua, Tonga. Bernice P. Bishop Mus. Bull. 96.

Jakes, P. and Gill, J. 1970. Rare earth elements and the island arc tholeiitic series. Earth Planet. Sci. Letters. 9, 17-28.

Jakes, P., and White, A.J.R. 1969. Structure of the Melanesian Arcs and correlation with distribution of magma types. Tectonophysics. 8, 223-236.

Jakes, P. and White, A.J.R. 1970. K/Rb ratios of rocks from island arcs. Geochim. Cosmochim. Acta. 34, 849.

Karig, D.E. 1971. Origin and development of marginal basins in the western Pacific. J. Geophys. Res. 76, 2542-2561.

Karig, D.E. and Mammerickx, J. (in press). Tectonic framework of the New Hebrides Island arc.

Kuno, H. 1959. Origin of Cenozoic petrographic provinces of Japan and surrounding areas. Bull. Volcanol. 20, 37-76.

Kuno, H. 1966. Lateral variation of basalt magma across continental margins and island arcs. Bull. Volcanol. 29, 195-222.

Ladd, H.S., and Hoffmeister, J.E. 1945. Geology of Lau, Fiji. Bernice P. Bishop Mus. Bull. 181

LePichon, X., and Heirtzler, J.R. 1968. Magnetic anomalies in the Indian Ocean and sea-floor spreading. J. Geophys. Res. 73, 2101-2117.

Lillie, A.R., and Brothers, R.N. 1970. The geology of New Caledonia. N.Z. J. Geol. Geophys. 13, 145-183.

Malahoff, A. (preprint). The Fiji Plateau: an active spiralling tectonic system.

Mallick, D.I.J. (this volume). The New Hebrides Archipelago.

McDougall, I. 1963. K-Ar Ages of Some Rocks from Viti Levu, Fiji. Nature. 198, 677.

Melson, W.G., Jarosewhich, E., and Lundquist, C.A. 1970. Volcanic eruption at Metis Shoal, Tonga, 1967-1968: description and petrology. Smithsonian Contr. to Earth Sci. No. 4.

Mitchell, A.H.G. 1966. Geology of South Malekula. New Hebrides Geol. Surv. Rep. 3.

Mitchell, A.H. and Reading, H.G. 1971. Evolution of island arcs. J. Geol. 79, 253-284.

Mitchell, A.H.G., and Warden, A.J. 1971. Geological evolution of the New Hebrides island arc. Jour. Geol. Soc. London.

Moberly, R. (this volume). Youthful oceanic lithosphere of marginal seas, western Pacific.

Page, R.W. 1971. The geochronology of igneous rocks in the New Guinea region. Unpub. Ph.D. thesis, A.N.U.

Page, R.W., and McDougall, I. 1970. Potassium-argon dating of the Tertiary f_{1-2} stage in New Guinea and its bearing on the geological time-scale. Am. J. Sci. 269, 321-342.

Phillips, K.A. 1965. Provisional Geological Map of Fiji. Govt. Printers: Suva, Fiji.

Richard, J.J. 1962. Catalogue of the active volcanoes of the world. Part 13, Kermadec, Tonga and Somoa.

Rickard, M.J. 1966. Reconnaissance geology of Vanua Levu. Fiji. Geol. Surv. Mem. 2.

Robinson, G.P. 1969. The geology of North Santo. New Hebrides Geol. Surv. Rept.

Rodda, P. 1967. Outline of the geology of Viti Levu. N.Z. J. Geol. and Geophys. 10, 1260-1273.

Rodda, P. 1969. Analyses of rocks from Fiji. Geol. Surv. Dept., Fiji.

Rodda, P., Snelling, N.J., Rex, D.C. 1967. Radiometric age data on rocks from Viti Levu, Fiji. N.Z. J. Geol. Geophys. 10, 1248-1259.

Santo, T. 1970. Regional study of the characteristic seismicity of the world. Part III: New Hebrides region. Bull. Earthquakes Res. Inst. 48, 1-18.

Sclater, J.G., Hawkins, J.W., and Mammerickx, J. (in press). Crustal extension between the Tonga and Lau Ridges: petrologic and geophysical evidence.

Suess, E. 1904. The Face of the Earth. Vol. 4. Oxford: Clarendon Press.

Sugimura, A. 1968. Spatial relations of basaltic magmas in island arcs. In: Basalts (eds. H.H. Hess and A. Poldevaart) V.2, 537-571.

Taylor, S.R., Capp, A.C., Graham, A.L., and Blake, D.H., 1969. Trace element abundances in andesites II Saipan, Bougainville and Fiji. Contr. Mineral. and Petrol. 23, 1-26.

Tomita, T. 1935. On the chemical composition of the Cenozoic alkaline suite of the circum-Japan-Sea region. Shanghai Sci. Inst. Jour. I, 227-306.

Warden, A.J. 1967. The geology of the Central Islands. New Hebrides Geol. Surv. Rep. 5.

Williams, C.E.F. and Warden, A.J. 1964. Progress report of the Geological Survey, 1959-1962. New Hebrides Geol. Surv. Rep.

STRONTIUM ISOTOPES IN ISLAND ARC VOLCANIC ROCKS

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ABSTRACT

The 87Sr/86Sr ratios of island arc volcanic rocks rule out both old sialic crust and unmodified ocean floor basalts as principal source materials for island arc volcanism.

There is a positive correlation between Rb/Sr and 87Sr/86Sr ratios in some island arc tholeiitic suites and a decrease of 87Sr/86Sr ratios with increasing distance from the trench in some island arcs. Although alteration and/or metamorphism of ocean floor basalts could produce the requisite Sr isotope ratios prior to subduction, the non-randomness of ratio variations in island arcs may discount this possibility. Instead, these variations suggest either that sediments contribute to magmas formed during subduction and that this contribution decreases with depth, or that the cold down-going slab is isotopically unequilibrated before fusion but that equilibration increases as depth and temperatures increase.

Because decreasing 87Sr/86Sr ratios accompany increasing K, Rb and Sr contents and Rb/Sr ratios in some island arc volcanic rocks, these variations in alkali contents and ratios cannot be explained by inferring additional crustal contamination or mantle wall-rock reaction as depth to the Benioff Zone increases.

1. Introduction

The isotope ratio 87Sr/86Sr can be used to identify Sr which has resided in the sialic crust. This is possible because most of the Earth's Rb lies in the crust whereas most of its Sr is in the mantle (Hedge and Walthall, 1963). Because the isotope 87Rb changes to 87Sr by radioactive decay, the amount of radiogenic Sr and the ratio of radiogenic to stable Sr (i.e. 87Sr/86Sr) will increase more quickly in the crust than mantle.

A typical Rb-Sr inventory for the Earth is given in Table 2 of Armstrong (1968). The 87Sr/86Sr ratio of the Earth at the time of its formation was probably close to 0.6990 (a value obtained from studies of meteoritic and lunar samples; cf. Papanastassiou and Wasserburg 1969). Changes of this ratio in the sialic crust and mantle with time will depend on when and how they separated from one another. If the separation was gradual but largely accomplished within the first 10^9 years of Earth history, the development will be as is sketched in Figure 1. There we use the bulk Rb/Sr ratio given for the Earth by Armstrong and refer to this starting material as "undepleted mantle", capable of being fractionated to yield a "crustal component" and residual "depleted mantle". Modern Sr from the sialic crust will have an 87/86 ratio of about 0.715-0.72; Sr from undepleted mantle will have an 87/86 ratio of about 0.706; and Sr from mantle having lost some portion of its crustal component will have an 87/86 ratio between 0.701 and 0.706.

Details of such Sr isotopic evolution are much debated (Armstrong 1968; Hurley and Rand 1969; Hart and Brooks 1970) but it is undisputed that Sr from the sialic crust has been clearly labelled during Earth history.

Modern Sr isotope analyses usually have a precision of about $\pm$ 0.0001-5. All analytical data presented or discussed in this paper have been normalized so that the measured 87Sr/86Sr ratio for Eimer and Amend $SrCO_3$ would be 0.7080 in any laboratory from which data have been used.

II Sr from the modern mantle

Recent intra-oceanic volcanism is most likely to sample the 87Sr/86Sr ratio of the modern mantle. Such volcanism produces rocks having ratios between 0.701 and 0.706 (Gast 1967; Peterman and Hedge 1971). Fresh ridge tholeiites from the Atlantic, Pacific, and Indian Oceans have a mean ratio of 0.7025; the range is from 0.7020 to 0.7030 except for one sample at 0.7012 (Hedge and Peterman 1970; Hart 1971). In contrast, tholeiitic and alkali basalts of oceanic islands have ratios of 0.703 to 0.706. Rb/Sr ratios likewise differ, being ~0.008 in ridge tholeiites but >0.02 in island basalts.

These different populations are illustrated in Figure 2 where it is apparent, as Tatsumoto et al. noted in 1965, that most ridge and some Hawaiian basalts have Rb/Sr ratios too low to produce the isotopic composition of their Sr even given the age of the Earth in which to do so. At the degree of partial melting thought necessary to produce tholeiitic basalt from a peridotitic source, Rb/Sr ratios of the melt will approximate those of the source (Griffin and Murthy, 1969). Thus

basalts whose analyses lie above the single-stage evolution line in Figure 2 (see caption for explanation) must have come from a source region which once had a higher Rb/Sr ratio than it did at the time of magma genesis (Gast 1967, 1968). This is thought to indicate widespread extraction of Rb (and K, Ba, La and Th) from the mantle at some time after Earth formation, probably associated with crustal development, leaving behind a mantle with heterogeneous Rb and Sr distributions. Peterman and Hedge (1971) take the ratio $K_2O/(Na_2O + K_2O)$ as a primary feature of magma compositions, little affected by fractionation. They note a positive correlation between 87Sr/86Sr and $K_2O/(Na_2O + K_2O)$ ratios in oceanic basalts and suggest that this indicates varying degrees of previous Rb and K extraction.

If these variations are due to random mantle heterogeneities, it is not clear why rocks with low ratios are limited to ocean ridges. Green (1971) suggests that alkali and large cation enrichment (and therefore higher Rb/Sr) occurs at the top of the low velocity zone (LVZ), and that ocean ridge volcanism taps mantle from the base of the LVZ which is depleted in these elements. However, if the Rb/Sr ratios of Hawaiian and ridge tholeiites reflect those of the upper and lower LVZ respectively, those differences must have existed for 1.5×10^9 years to increase 87Sr/86Sr ratios from 0.7025 to 0.7040.

If mantle phases do not achieve isotopic equilibrium on the grain-to-grain scale, as is discussed for the anhydrous moon by Compston and Gray (1972) and for alpine peridotites by Kuroda et al. (1971) and Graham and Ringwood (1971), one might attribute the high 87Sr/86Sr ratios of alkali basalts to small

degrees of partial melting involving phlogopite. Using the data of Griffin and Murthy (1969, Table 5), the 87/86 ratio of Sr in phlogopite would increase by 0.0015 in 2.6×10^6 years. However, isotopic disequilibrium on this scale is both difficult to imagine in a hydrous, high temperature environment, and does not explain the differences in isotopic composition of Sr between Hawaiian and ridge tholeiites.

Whatever the reason, Sr produced directly from the mantle can have 87/86 ratios between 0.701 and 0.706.

III Sr in western Pacific island arcs

Hedge (1966), Pushkar (1968), Kurasawa (1970), Hedge and Knight (1969), Ewart and Stipp (1968), Peterman et al (1970) and Gill (1970) have published data on the isotopic composition of Sr from western Pacific island arcs. New data for samples from New Georgia (Solomon Islands), the New Hebrides, and Fiji are given in Table 1 and Figure 3; data for Tonga, also presented in Figure 3, will be published separately. Table 2 summarizes the data available for the southwestern Pacific.

Figure 3 and Table 2 illustrate the consistency of island arc 87Sr/86Sr ratios, which range from 0.703 to 0.705. Note the positive correlation between Rb/Sr and 87Sr/86Sr in volcanics from Tonga and some of those from Fiji.

Many geochemical aspects of island arc magmas change as the distance from their associated trench and Benioff Zone increases (Sugimura 1968, this volume; Gill and Gorton, this volume; Jakes and Gill 1970; Dickinson and Hatherton 1967). Volcanics erupted closest to the trench are usually members of an island arc tholeiitic series, those further away belong to a calc-alkaline series, and those still further away represent a high-K variety of the calc-alkaline series, or a shoshonite

association (Jakes and Gill 1970; Gill and Gorton, this volume, Figure 1). Hedge and Knight (1969) report that 87Sr/86Sr ratios in volcanics from northern Honshu, Japan, decrease from 0.7041-0.7043 closest to the trench to 0.7026-0.7031 furthest from it. Data are not sufficient to assess variations with Rb/Sr within suites. (Kurasawa (1970) reports the opposite trend in southwest Honshu but this is due to high ratios in the alkali basalts of southwest Japan. These basalts differ in such distinctive features as K/Na ratios and TiO_2 contents from other island arc alkalic rocks and may only indirectly reflect the subduction process.)

The 87Sr/86Sr ratios of Late Miocene volcanic rocks from Fiji (Figure 3b) also decrease as distance from the inferred trench position increases. Rb/Sr ratios are more variable within than between suites over this distance. Gill and Gorton (this volume) argue on independent geochemical and geological grounds that Vanua Levu represents the Late Miocene volcanic front in Fiji, that the calc-alkaline andesites of southeast Fiji were erupted behind that front, and the high-K andesites of Kandavu still further behind it. If these arguments are sound and volcanics of these three islands were contemporaneous and related to the same subduction configuration, then we have documented a second case in which 87Sr/86Sr ratios decrease as the K,Rb, and Sr contents of the magmas increase.

Page (personal communication, 1971) has found no change in 87Sr/86Sr ratios between the north coast volcanics of New Britain and those of Talasea studied by him and Peterman et al. (1970). Stipp (1968) found a westward decrease in 87/86 ratios of basalt Sr in New Zealand.

IV Implications for magma genesis in island arcs

Island arc magmas have too little 87Sr to involve reworked sialic crust. Thus old, fractionated (high Rb/Sr) crust can be neither the source nor a major contaminant of island arc volcanism.

Yet island arc magmas have too much 87Sr to be simple partial melts of oceanic basalts having 87Sr/86Sr ratios of 0.7025 and thrust beneath island arcs along Benioff zones, even if these basalts are 10^8 years old. Nevertheless the number of alternative hypotheses consistent with the island arc data is legion.

The observed ratios are well within the range of mantle-derived volcanics and could therefore represent either various partial melts of hydrous peridotite (Kushiro et al. 1968) or fractionated derivatives of mantle-derived basalts. Neither, however, explain the instances of positive correlation between 87Sr/86Sr and Rb/Sr ratios.

Alternatively, underthrust basalts could be the source of island arc volcanism if their ratios were 0.703-0.705. This could be accomplished in at least two ways: partial equilibration of ocean floor basalt Sr with seawater Sr (87/86 = 0.709) during or after eruption; or isotopic disequilibrium within the downgoing slab. Hart and Nalwalk (1970) attribute ratios of 0.705-0.706 in dredged oceanic basalts to partial equilibration with seawater. Indirect equilibration might also occur during burial metamorphism. Gill (1970) reported high 87Sr/86Sr ratios

in some of his first period volcanics which had experienced burial metamorphism. In one case (sample 68-60) we have found the 87Sr to reside in an HCl-soluble phase so that Sr in the insoluble residue had an 87/86 ratio lower by 0.0011. An alternative or complementary suggestion is for the Rb/Sr ratio of ocean floor basalt to increase by alteration, leading eventually to increased 87Sr/86Sr ratios. Hart (1971, Figure 1) shows that Rb/Sr ratios can increase 2-7 times as a result of alteration, thus reducing the time necessary for development of 87Sr by an equivalent amount. Fusion of such altered and/or metamorphosed basalt could yield island arc volcanics with their observed 87Sr/86Sr ratios but the consistency of ratios in various island arcs requires remarkable uniformity for these alteration effects. Moreover, if isotopic equilibrium is attained within the oceanic crust, varying degrees of partial fusion would not produce the Rb/Sr-87Sr/86Sr correlation or the regular spatial variations in 87Sr/86Sr ratios discussed above.

A third alternative involving some kind of mixing process must be invoked to explain these variations. Authigenic oceanic sediments will have the 87Sr/86Sr ratios of sea water; detrital sediments will have ratios very similar to those of their provenence (Dasch 1969; Biscaye and Dasch 1971). Involvement of either during magma genesis would increase both 87Sr/86Sr and Rb/Sr ratios of the melt by an amount proportional to its weight fraction.

Many authors have appealed to sediment contamination as a way to produce the concentrations of alkali elements and radiogenic Sr and Pb in island arc volcanics (Armstrong 1968, 1971; Tatsumoto

1969; Jakes and Gill 1970; Hart et al 1970). Armstrong calculates that over ninety per cent of the lead in arc magmas is from oceanic sediments. That this is unlikely for Sr can be illustrated by referring to sample 68-66 in Table 1.

This andesite has been studied experimentally by Green (1972, and this volume) who suggests it represents a partial melt derived from eclogite. Figure 4 illustrates the Sr contents to be expected from melting an eclogite of oceanic tholeiite composition containing 110 ppm Sr and with a clinopyroxene + garnet + rutile residuum. We have used the equations of Shaw (1970) and taken $K_{Sr}^{liq/cpx}=6.7$, $K_{Sr}^{cpx/gar}=20.$, and $K_{Sr}^{liq/rut}=\infty$ (Gast 1968; Philpotts and Schnetzler 1970). As discussed by Gill (in preparation) a 19% partial melt of oceanic tholeiitic best approximates the major element composition of sample 68-66, using Green's data for the composition of liquidus clinopyroxene and garnet at experimental conditions of 36 kb, 1100°C, and 10% H_2O. From Figure 4 this should yield Sr concentrations between 430-530 ppm in the melt depending on whether or not the liquid and residuum have maintained equilibrium during melting. Sample 68-66 has 481 ppm Sr and an 87Sr/86Sr ratio of 0.7040 (Gill 1970).

The amount of sediment Sr needed to increase the 87/86 ratio of a melt depends on the 87Sr/86Sr ratio of the sediment. The maximum ratio to be expected would be 0.72. A more realistic estimate for western Pacific sediments undergoing subduction would lie between the ratio for sea water (0.709) and that of volcaniclastic greywacke (e.g. 0.7066; Ewart and Stipp 1968). To increase the 87Sr/86Sr ratio from 0.7025 (ocean floor basalt) to 0.7040 (sample 68-66) requires assimilation of 41 ppm Sr if its 87/86 ratio is 0.72, 124 ppm Sr if its 87/86 ratio is 0.709, and 198 ppm Sr if its 87/86 ratio is 0.7066. The agreement between the Sr content of sample 68-66 and that predicted

from melting eclogite alone is sufficiently close that sediment assimilation is unlikely to provide the sole or even major explanation for its 87/86 ratio. Deep sea sediments are more likely to be clays than carbonates, and such clays will have approximately 180 ppm Sr (Turekian and Wedepohl 1961). Appeal to the incorporation of sediments in order to explain the 87/86 ratio of Sr in sample 68-66 requires melting more sediment than basalt during magma genesis, a possibility excluded by many other geochemical parameters (Gill, in preparation).

A different kind of mixing process could occur if isotopic equilibration is impeded by the low temperatures within the slab. If so, phases with high Rb/Sr and therefore eventually high 87Sr contents might persist until melting. In a hydrous environment, initial melts from basalt will be quartz-rich and will presumably extract elements like Rb (Green, 1972 and this volume). In Tonga and Fiji, rhyodacites (SiO_2 = 63-80%) of the island arc tholeiitic series have higher Rb/Sr and 87Sr/86Sr ratios than do associated andesites and basalts, and may reflect such a situation.

It may be significant that this correlation is most pronounced in rocks of the island arc tholeiitic series which are associated with the shallowest earthquake foci and therefore below which the slab is most likely to remain hydrous and cold, and its basalts to exist in amphibolite facies assemblages. The rhyodacites from Fiji with higher Rb/Sr and 87Sr/86Sr ratios contain no more Sr than do oceanic tholeiites themselves and this requires $K_{Sr}^{liq/xl}\sim 1.0$ during magma genesis. Such a bulk distribution coefficient is possible in an amphibolite facies residuum composed of amphibole, clinopyroxene, and plagioclase but not in an eclogite facies mixture of clinopyroxene and

and garnet, using the data of Gast (1968), Philpotts and Schnetzler (1970) and Korringa and Noble (1970).

In situations where 87Sr/86Sr ratios of volcanic rocks decrease as distance between these rocks and the trench or Benioff zone increases, simple melting of any isotopically homogeneous source is ruled out. If the downgoing slab is that source, the amount of 87Sr from it which enters the melt must decrease with depth. If the 87Sr comes from sediments the amount of sediment contaminating any magmas produced must also decrease with depth. The opposite conclusion was reached by Donnelly et al. (1971) using relative alkali element abundances. Higher concentrations of these elements more likely indicate smaller degrees of melting than additional sediment contamination. Alternatively, if the 87Sr comes from a high Rb/Sr phase whose Sr remains isotopically unequilibrated with its surrounding, either that phase must disappear during dehydration or isotopic equilibrium must be approached as depth and temperatures increase. If overlying mantle rather than the slab is the source region for island arc volcanism, Rb/Sr ratios of that mantle are regularly zoned either with depth or distance from the trench. Either or both could be consistent with earlier fractionation of Rb from Sr during crustal growth.

Unlike variations in K and other alkali contents (Dickinson and Hatherton, 1967; Gill and Gorton, this volume) a decrease in 87Sr/86Sr ratios with distance from the trench cannot be explained by increased crustal contamination or mantle "wall-rock reaction" as depth to the Benioff Zone increases. It must be a primary magma feature but does not discriminate between melts from the mantle or underthrust lithosphere.

V Conclusion

The average 87Sr/86Sr ratio in island arc volcanic rocks rules out both sialic crust and unmodified oceanic tholeiite as their principal source material. There are, nevertheless, so many alternatives by which to explain this average ratio that routine Sr isotope measurements are unlikely to make further positive contributions to the study of island arc magma genesis. Narrowing the list of alternatives requires a realistic assessment of Sr contents to be expected in partial melts, accurate identification of small (<0.001) variations in 87Sr/86Sr ratios within island arcs, and faith that nature is uniform on that scale.

TABLE 1

New Sr isotope data from Melanesia*

	Sample No.	Rb	Sr	87Sr/86Sr
A. FIJI (Late Miocene)				
1. Vanua Levu				
	798	8.8	202.	.7041
	806	14.	415.	.7041
	810	6.2	169.	.7040
	814	9.1	737	.7041
	818	16.	393.	.7035
	820	6.1	432.	.7037
	821	16.	347.	.7037
	M96	6.1	95.	.7041
	B535	12.	92.	.7043
2. Viti Levu (Namosi Andesite)	66	16.	481.	.7040
	873	35.	536.	.7036
	874	28.	652.	.7035
	876	17.	536.	.7037
3. Kandavu	390	51.	1330.	.7032
B. FIJI (Late Eocene to Early Miocene; also see Gill (1970)				
	60	2.1	128.	.7037
	387	3.1	90.	.7045
	903	31.	256.	.7034
	904	8.8	353.	.7032
	915	3.3	160.	.7041
	920	5.1	279.	.7032
C. NEW HEBRIDES (Aoba)				
	576	17.	643.	.7039
D. SOLOMONS (New Georgia)				
	364	3.9	90.	.7040
	414	2.8	110.	.7035
	372	6.1	80.	.7039
	450	10.	200.	.7040
	443	13.	140.	.7030
	413	13.	140.	.7039
	412	75.	470.	.7036
	536	18.	250.	.7043
	377	22.	210.	.7040
	428	23.	200.	.7037
	230	22.	290.	.7038
	325	24.	300.	.7035
	349	21.	170.	.7038

* For chemical composition and description of samples analyzed see Hindle and Gill (in prep) for Vanua Levu; Gill (1970 and in prep.) for other Fiji samples; and Stanton and Bell (1969) for New Georgia. All isotope ratios have been normalized to 8.37520 for 88Sr/86Sr and 0.7080 for E&A $SrCO_3$ 87Sr/86Sr. Replicate analyses of E&A $SrCO_3$ and rock samples consistently have a precision of $\pm$0.0001-2. Rb and Sr contents were determined by XRF with a precision of $\pm$ 5% for the lowest Rb concentrations and $\pm$ 2% of Sr for Fiji samples. New Georgia Rb and Sr contents were determined by a less precise XRF method and are approximate only. Their Rb/Sr ratios should be correct within $\pm$ 10%.

TABLE 2 Summary of Sr isotope compositions in Southwest Pacific volcanic rocks.

AREA	87Sr/86Sr * Mean	Range	SOURCE
1. Talasea, New Britain	.7035	.7034 - .7038	Peterman et al (1970)
2. New Georgia, Solomon Islands	.7037	.7030 - .7042	this paper
3. Aoba, New Hebrides	.7039		"
4. Vanua Levu, Fiji	.7040	.7036 - .7043	"
5. Viti Levu, Fiji	.7038	.7032 - .7048	" and Gill (unpub)
6. Kandavu, Fiji	.7032		this paper
7. Lau, Fiji	.7032	.7030 - .7033	Gill (unpub)
8. Tonga	.7040	.7038 - .7043	this paper
9. Taupo, New Zealand	.7054	.7040 - .7066	Ewart and Stipp (1968)

* All ratios normalized to 0.7080 for E. and A. $SrCO_3$.

FIGURE CAPTIONS

FIGURE 1. Evolution of 87Sr/86Sr in various environments during Earth history.

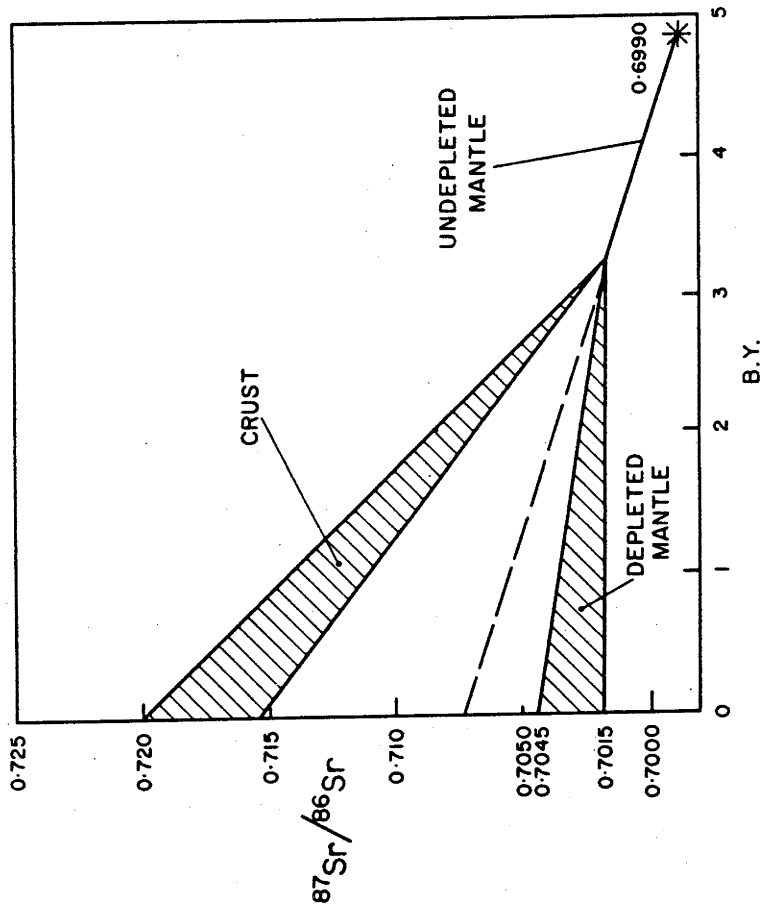
FIGURE 2. Variation of 87Sr/86Sr with Rb/Sr ratios in oceanic basalts for which this information is available. Data is from Bence (1966), Tatsumoto et al. (1965), and Hart (1971) for ocean floor basalts. Open circles represent data on Hawaiian basalts taken from Powell and DeLong (1966).

The line terminating at "C+M" is a single stage evolution line which defines the 87Sr/86Sr ratio of all environments whose Rb/Sr ratios have remained constant throughout 4.6×10^9 years of Earth history. All points above this line represent basalts from source regions whose Rb/Sr ratios were too low at the time of partial fusion to have generated sufficient 87Sr from 87Rb during that time to explain the observed 87Sr/86Sr ratios of those basalts. Crystal fractionation usually leads to higher Rb/Sr ratios but should not affect 87Sr/86Sr ratios.

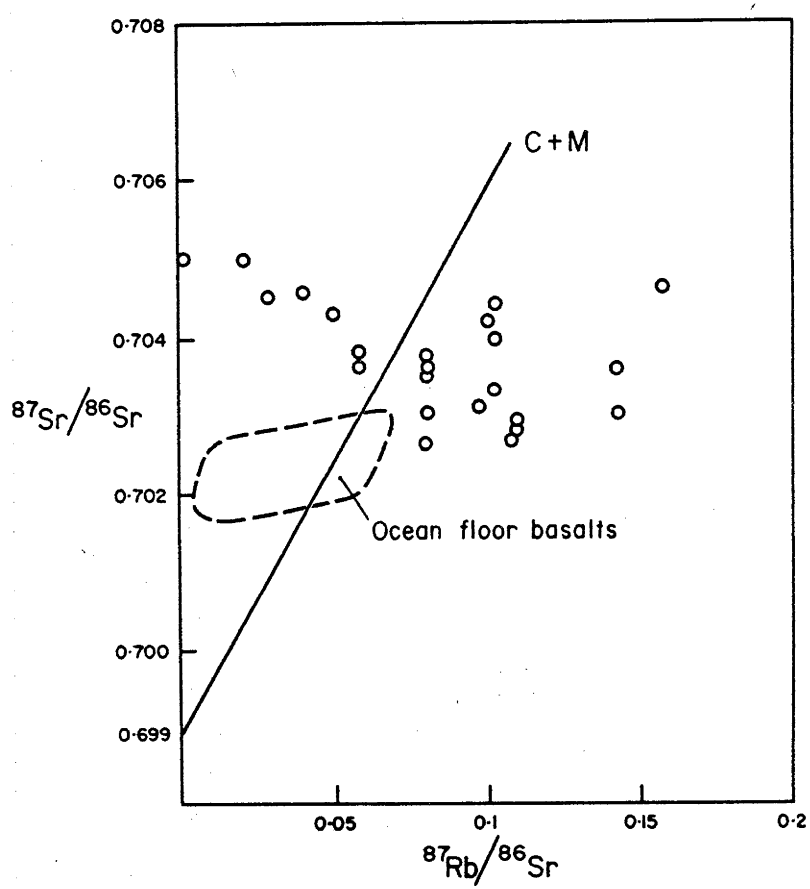
FIGURE 3 Variation of 87Sr/86Sr with Rb/Sr ratios in Melanesian volcanic rocks. Data is from Table 1 and Gill (1970 and unpublished).

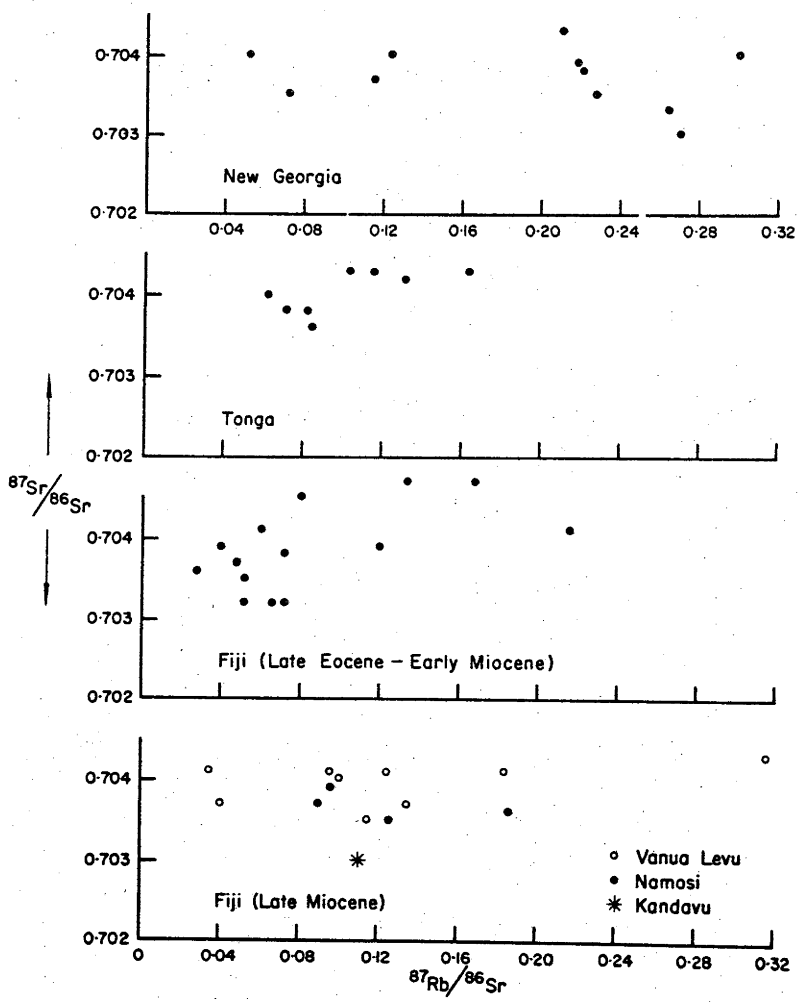
FIGURE 4 Predicted concentration of Sr in partial melts of eclogite.

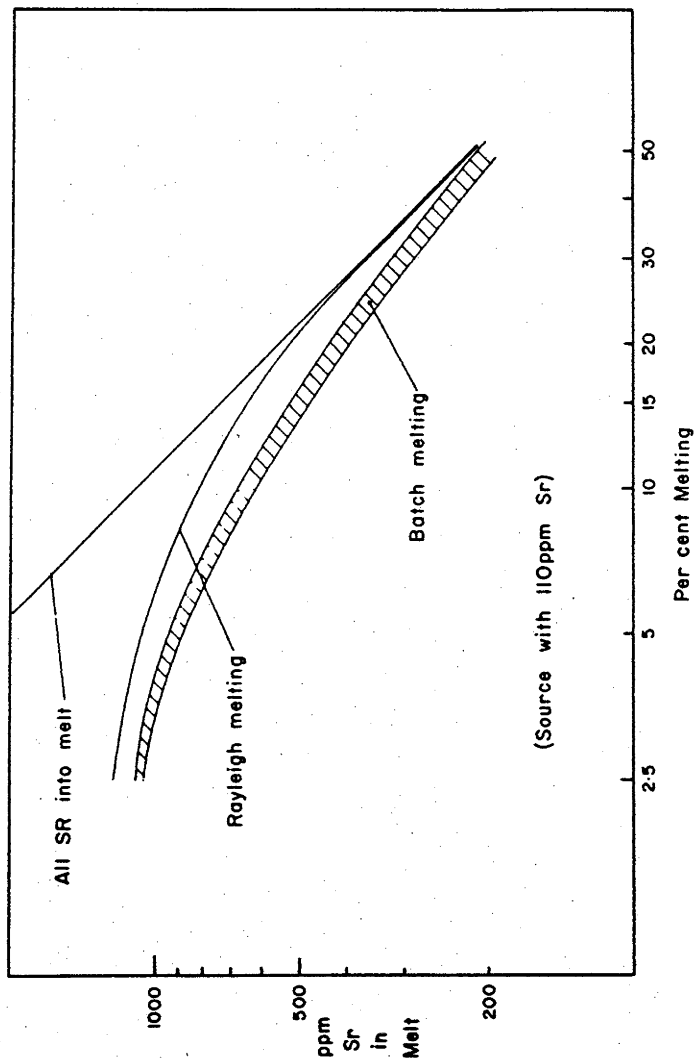
1


UNDEPLETED MANTLE
0·6990
CRUST
DEPLETED MANTLE
B.Y.
0
1
2
3
4
5
0·725
0·720
0·715
0·710
0·7050
0·7045
0·7015
0·7000
$^{87}Sr/^{86}Sr$

2


0·708
0·706
0·704
0·702
0·700
0·699
C+M
Ocean floor basalts
$^{87}Sr/^{86}Sr$
0·05
0·1
0·15
0·2
$^{87}Rb/^{86}Sr$


New Georgia
Tonga
Fiji (Late Eocene – Early Miocene)
Fiji (Late Miocene)
Vanua Levu
Namosi
Kandavu
$^{87}Sr/^{86}Sr$
$^{87}Rb/^{86}Sr$
0·704
0·703
0·702
0·04
0·08
0·12
0·16
0·20
0·24
0·28
0·32
0


All SR into melt
Rayleigh melting
Batch melting
(Source with 110ppm Sr)
Per cent Melting
50
30
20
15
10
5
2·5
1000
500
200
ppm Sr in Melt

REFERENCES

Armstrong, R.L., (1968). A model for the evolution of Strontium and Lead isotopes in a dynamic Earth. Rev. Geophys. 6, 175-199.

_____ (1971) Isotopic and chemical constraints on models of magma genesis in volcanic arcs. Earth Planet. Sci. Letters. 12, 137-142.

Bence, A.E., (1966) The differentiation history of the earth by rubidium-strontium isotopic relationships. In: Variations in isotopic abundances of strontium, calcium, and argon and related topics (ed. P.M. Hurley), M.I.T. 1381-14, 14th Ann. Prog. Rep., pp. 35-78.

Biscaye, P.E. and Dasch, E.J. (1971) The rubidium, strontium, strontium-isotope system in deep-sea sediments: Argentine Basin. J. Geophys. Res. 76, 5087-5096.

Compston, W. and Gray, C.M. (1972) Lunar model Rb-Sr ages and magma genesis. Earth Planet. Sci. Letters.

Dasch, E.J. (1969) Strontium isotopes in weathering profiles, deep-sea sediments, and sedimentary rocks. Geochim. Cosmochim. Acta 33, 1521-1552.

Dickinson, W.R. and Hatherton, T. (1967) Andesitic Volcanism and Seismicity around the Pacific. Science 157, 801-803.

Donnelly, T.W., Rogers, J.J.W., Pushkar, P., and Armstrong, R.L. (1971). Chemical evolution of the igneous rocks of the Eastern West Indies: an investigation of thorium, uranium, and potassium distributions, and lead and strontium isotopic ratios. G.S.A. Mem. 116.

Ewart, A. and Stipp, J.J. (1968) Petrogenesis of the volcanic rocks of the Central North Island, New Zealand, as indicated by a study of Sr 87/86 ratios, and Sr, Rb, K, U, and Th abundances. Geochim. Cosmochim. Acta 32, 699-736.

Gast, P.W. (1967) Isotope geochemistry of Volcanic Rocks. In: Basalts (eds. H. Hess and A. Poldevaart), vol. 1, 325-358.

Gast, P.W. (1968) Trace element fractionation and the origin of tholeiitic and alkaline magma types. Geochim. Cosmochim. Acta 32, 1057-1086.

Gill, J.B. (1970) Geochemistry of Viti Levu, Fiji, and its evolution as an island arc. Contr. Mineral Petrol. 27, 179-203.

Gill, J. and Gorton, M. (this volume). A proposed geological and geochemical history of Eastern Melanesia.

Graham A.L. and Ringwood A.E. (1971) Lunar-basalt genesis: the origin of the europium anomaly. Earth Planet. Sci. Letters

Green, D.H. (1971) Composition of basaltic magmas as indicators of conditions of origin: application to oceanic volcanism. Phil. Trans. Roy. Soc. Lond. A. 268, 707-725.

Green, T. (1972) Crystallization of calc-alkaline andesite under controlled high-pressure hydrous conditions. Contr. Mineral. and Petrol.

Griffin, W.L. and Murthy, V.R. (1969) Distribution of K, Rb, Sr and Ba in some minerals relevant to basalt genesis. Geochim. Cosmochim. Acta 33, 1389-1414.

Hart, S.R., (1971) K, Rb, Cs, Sr, and Ba contents and Sr isotope ratios of ocean floor basalts. Phil. Trans. Roy. Soc. Lond. A 268, 573-587.

Hart, S.R. and Brooks, C. (1970). Rb-Sr mantle evolution models. Dept. Terrestrial Magnetism Ann. Rep. 1968-1969, pp. 426-429.

Hart, S.R., Brooks, C., Krogh, T.E., Davis, G.L., and Nava, D. (1970). Ancient and modern volcanic rocks: a trace element model. Earth Planet. Sci. Letters 10, 17-28.

Hart, S.R. and Nalwalk, A.J. (1970). K, Rb, Cs and Sr relationships in submarine basalts from the Puerto Rico trench. Geochim. Cosmochim. Acta 34, 145-156.

Hedge, C.E. (1966) Variations in radiogenic strontium found in volcanic rocks. J. Geophys. Res. 71, 6119.

Hedge, C.E., and Knight, R.J., (1969). Lead and strontium isotopes in volcanic rocks from northern Honshu, Japan. Geochem. J. 3, 15-24.

Hedge, C.E. and Peterman, Z.E. (1970). The strontium isotopic composition of basalts from the Gordo and Juan de Iuca Rises, Northeastern Pacific Ocean. Contr. Mineral.and Petrol. 27, 114-120.

Hedge, C. and Walthall, F. (1963) Radiogenic Sr^{87} as an index of geologic processes. Science 140, 1214-1217.

Hurley, P.M. and Rand, R.J. (1969) Pre-drift continental nucleii. Science 164, 1229-1242.

Jakes, P. and Gill, J. (1970). Rare earth elements and the island arc tholeiitic series. Earth Planet. Sci. Letters 9, 17-28.

Korringa, M.K. and Noble, D.C. (1970) Feldspar-liquid partition coefficients for Sr and Ba based on a Rayleigh fractionation model. Abs. Geol. Soc. Am. Ann. Meeting, p. 598.

Kurasawa, H. (1970) Strontium and lead isotopes of volcanic rocks in Japan. In: Recent Developments in Mass Spectrometry (eds. K. Ogata and T. Haya kawa), pp. 666-670.

Kuroda, V., Iizumi S., and Tazaki, K. (1971) On some problems in origin of the Alpina type ultramafic rocks. Abs. Int'l. Geochem. Congr. (Moscow), vol. I, pp. 65-66

Kushiro, I., Yoder, H.S. and Nishikawa, M. (1968) Effect of water on the melting of enstatite. Geol. Soc. Am. Bull. 79, 1685-1692.

Papanastassiou, D.A. and Wasserburg, G.J. (1969) Initial strontium isotopic abundances and the resolution of small time differences in the formation of planetary objects. Earth Planet. Sci. Letters 5, 361-376.

Peterman, Z.E. and Hedge, C.E. (1971) Related strontium isotopic and chemical variations in oceanic basalts. Bull. Geol. Soc. Am. 82, 493-500.

Peterman, Z.E., Lowder, G.G., and Carmichael, I.S.E. Sr87/Sr86 ratios of the Talasea series, New Britain, Territory of New Guinea. Bull. Geol. Soc. Am. 81, 39-40.

Philpotts, J.A. and Schnetzler, C.C. (1970) Phenocryst-matrix partition coefficients for K, Rb, Sr and Ba, with applications to anorthosite and basalt genesis. Geochim. Cosmochim. Acta. 34, 307-322.

Powell, J.L. and DeLong S.E. (1966) Isotopic composition of strontium in volcanic rocks from Oahu. Science 153, 1239-1242.

Pushkar, Paul (1968) Strontium Isotope Ratios in Volcanic Rocks of Three Island Arc Areas. J. Geophys. Res. 73, 2701-2713.

Shaw, D.M. (1970) Trace element fractionation during anatexis. Geochim. Cosmochim. Acta 34, 237-242.

Stanton, R.L. and Bell, J.D. (1969) Volcanic and associated rocks of the New Georgia Group, British Solomon Islands Protectorate. Overseas Geol. Min. Resources 10, 113-145.

Stipp, J.J. (1968) The geochronology and petrogenesis of the Cenozoic volcanics of North Island, New Zealand. Unpub. Ph.D. thesis, A.N.U.

Sugimura, A. (1968). Spatial relations of basaltic magmas in island arcs. In: Basalts (eds. H.H. Hess and A. Poldevaart) v.2, 537-571.

Tatsumoto, M. (1969). Lead isotopes in volcanic rocks and possible ocean-floor thrusting beneath island arcs. Earth Planet. Sci. Letters 6, 369-376.

Tatsumoto, M., Hedge, C.E., and Engel A.E.J., (1965) Potassium, rubidium, strontium, thorium,uranium and the ratio of Sr-87 to Sr-86 in oceanic tholeiitic basalt. Science 150, 886-888.

Turekian, K.K. and Wedepohl, K.H. (1961) Distribution of the elements in some major units of the Earth's crust. Bull. Geol. Soc. Am. 72, 175-192.

Appendix 4. C.I.P.W. Norms of Rocks with < 52% SiO_2 (assuming $Fe_2O_3/FeO = 0.25$).

	61	62	63	64	65	71	72	73	74	75	76	367
qz	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
or	5.79	4.67	24.43	19.80	20.85	16.29	10.51	23.35	9.25	12.83	24.59	5.71
ab	23.98	25.15	14.34	24.28	22.98	19.26	13.59	27.87	18.11	21.23	28.06	23.65
an	36.13	37.64	12.83	32.24	31.05	22.57	17.45	20.72	15.94	30.73	29.23	35.57
ne	0.0	0.0	15.73	0.43	1.71	3.04	4.70	0.83	3.46	0.0	0.68	0.0
di	9.91	6.75	25.00	8.89	9.27	21.18	31.54	11.85	32.21	17.17	6.16	11.78
hyp	1.96	7.60	0.0	0.0	0.0	0.0	0.0	0.0	0.0	3.75	0.0	10.58
ol	16.09	11.82	3.50	10.02	9.75	11.87	17.34	10.15	16.14	9.73	7.20	6.96
mag	3.25	3.36	2.02	2.08	2.10	2.96	2.93	2.49	2.67	2.53	1.80	2.99
ilm	2.16	2.25	1.06	1.20	1.20	1.93	1.12	1.39	1.29	1.42	1.05	2.05
ap	0.72	0.76	1.08	1.06	1.09	0.92	0.82	1.34	0.93	0.61	1.24	0.73

	368	372	376	378	382	389	394	795	799	806	807	808
qz	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.31	0.84	0.0	0.0
or	4.44	3.67	6.94	4.62	3.77	11.56	9.46	4.84	3.58	4.64	5.74	3.99
ab	19.14	16.35	29.64	17.00	16.29	20.20	15.75	26.05	18.04	22.39	26.24	25.39
an	41.75	39.92	26.21	39.25	28.28	30.24	23.09	29.97	28.65	29.15	33.50	31.09
ne	0.0	0.0	0.80	0.0	0.0	0.63	0.0	0.0	0.0	0.0	0.0	0.0
di	13.41	16.01	10.87	14.90	22.72	22.35	25.52	16.05	20.31	17.06	13.76	16.95
hyp	9.38	12.35	0.0	8.91	7.89	0.0	3.02	8.85	24.82	19.82	5.25	10.06
ol	6.73	6.83	16.45	10.07	16.19	9.75	18.12	6.08	0.0	0.0	7.46	4.37
mag	2.88	2.82	3.20	3.12	2.74	2.67	3.00	2.97	2.61	2.80	2.80	2.94
ilm	1.71	1.60	4.70	1.68	1.33	1.76	1.34	4.31	1.33	2.66	4.28	4.34
ap	0.55	0.46	1.19	0.46	0.79	0.85	0.69	0.88	0.35	0.64	0.98	0.87

	816	818	821	824	825	826	827	828	829	830	831	840
qz	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.99	0.0
or	5.14	3.54	3.28	6.49	5.83	5.57	5.24	6.57	5.66	5.39	3.66	8.12
ab	16.61	18.08	17.57	22.29	28.77	31.08	19.70	25.03	23.36	27.45	20.19	17.36
an	29.98	31.43	26.55	27.96	29.80	32.77	29.31	29.17	28.53	33.69	39.16	25.94
ne	0.0	0.0	0.0	2.34	1.80	0.80	2.30	0.98	2.75	0.98	0.0	0.17
di	22.10	23.73	22.78	17.93	15.08	11.14	18.97	15.37	15.64	15.48	8.50	28.82
hyp	13.35	17.86	13.94	0.0	0.0	0.0	0.0	0.0	0.0	0.0	21.32	0.0
ol	8.33	0.46	11.56	15.84	11.36	12.31	17.70	15.19	16.95	10.11	0.0	14.47
mag	2.59	3.13	2.60	2.73	2.78	2.68	2.74	2.80	2.57	2.40	2.90	2.88
ilm	1.55	1.41	1.13	3.37	3.44	3.29	3.10	3.46	3.51	3.49	1.77	1.65
ap	0.35	0.37	0.41	1.05	1.16	1.08	0.93	1.26	1.03	1.01	0.51	0.59

	841	842	843	850	853	855	860	861	862	863	864	865
qz	5.49	0.0	1.52	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
or	4.39	3.26	6.92	4.50	0.74	2.12	7.24	21.25	6.17	22.38	23.27	16.76
ab	26.12	16.97	35.62	39.38	22.47	25.57	27.02	8.58	14.03	19.44	26.11	17.90
an	37.32	33.23	27.12	31.96	39.12	38.42	34.07	17.93	26.76	19.11	32.69	20.47
ne	0.0	0.64	0.0	0.0	0.0	0.0	0.0	11.64	10.95	6.54	0.0	3.51
di	3.38	25.26	11.06	7.10	13.69	14.94	11.48	26.33	29.96	19.75	1.59	27.08
hyp	18.21	0.0	13.05	2.70	18.19	9.00	7.32	0.0	0.0	0.0	2.12	0.0
ol	0.0	14.66	0.0	9.95	0.91	5.02	7.26	8.50	6.38	8.09	9.54	8.30
mag	2.51	3.32	2.33	2.70	3.18	2.68	2.75	2.78	2.91	2.34	1.97	2.76
ilm	1.45	2.15	1.48	1.51	1.53	1.89	2.41	1.40	2.30	1.25	1.28	1.32
ap	1.12	0.52	0.88	0.20	0.17	0.38	0.45	1.59	0.55	1.08	1.43	1.90

Appendix 4 continued...

	881	882	884	886	887	888	893	894	895	896a	897	898
qz	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
or	24.79	23.09	19.44	15.10	11.49	17.58	13.59	35.99	11.59	14.49	22.25	25.21
ab	12.87	19.14	24.61	27.12	28.59	31.18	10.80	30.32	13.45	23.89	14.79	11.08
an	21.04	23.24	32.56	28.57	29.70	23.15	21.21	19.83	19.81	25.38	24.44	22.22
ne	11.07	6.91	0.37	0.0	0.0	3.61	0.64	0.0	1.94	0.0	2.23	6.71
di	17.83	16.30	6.85	10.36	14.30	9.58	25.82	1.78	30.89	16.53	16.98	20.13
hyp	0.0	0.0	0.0	13.56	10.90	0.0	0.0	1.91	0.0	9.23	0.0	0.0
ol	7.57	7.01	11.52	0.77	0.17	9.86	23.38	6.45	17.56	5.09	13.37	9.14
mag	2.24	2.11	2.22	2.32	2.36	2.28	2.77	1.62	2.88	2.88	2.85	2.74
ilm	1.37	1.13	1.36	1.33	1.44	1.25	1.01	1.07	1.10	1.66	1.44	1.30
ap	1.24	1.06	1.08	0.85	1.06	1.51	0.79	1.04	0.78	0.85	1.64	1.46

	899	900	902	905	907	911	912	916	917	918	919	920
qz	0.0	0.0	0.0	0.0	0.0	2.74	0.0	0.0	0.0	0.0	0.0	1.61
or	14.71	10.84	12.45	2.38	11.33	7.03	10.14	5.15	5.31	26.94	14.63	2.60
ab	20.63	16.46	4.87	35.54	22.76	25.30	21.63	20.83	21.50	19.13	8.38	19.77
an	23.89	22.65	12.69	26.18	23.16	35.29	27.75	35.82	36.00	23.08	17.16	38.82
ne	0.19	0.77	7.84	0.28	0.93	0.0	1.07	0.0	0.0	3.25	4.70	0.0
di	21.69	26.56	36.48	7.24	21.96	9.83	17.86	14.65	14.09	11.31	35.24	14.69
hyp	0.0	0.0	0.0	0.0	0.0	15.39	0.0	14.85	13.46	0.0	0.0	16.96
ol	13.73	17.95	20.28	21.38	13.72	0.0	15.41	4.08	5.02	10.89	14.51	0.0
mag	2.68	2.59	3.03	3.54	3.11	2.49	2.91	2.69	2.63	2.46	3.17	3.12
ilm	1.55	1.34	1.19	3.02	2.09	1.45	2.20	1.42	1.45	1.32	1.35	2.04
ap	0.93	0.83	1.17	0.45	0.96	0.48	1.03	0.51	0.54	1.62	0.87	0.39

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Geochemistry of Viti Levu, Fiji, and Its Evolution as an Island Arc

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Abstract. Available geochemical and field data suggest that three different periods of igneous activity have occurred on Viti Levu. Rocks of the first period are "island arc tholeiites"; of the second, a calc-alkaline series; and of the third, shoshonites. Each period has a geochemical distinctiveness. The temporal sequence also corresponds to a spatial zonation analogous to that observed in many currently active island arcs. K_2O and the larger trace elements increase south to north and with time, whereas iron enrichment and Na_2O/K_2O ratios decrease. Initial Sr 87/86 ratios average about 0.7041 in rocks of all periods. If Viti Levu magmas were genetically related to underthrust oceanic crust, as is argued for comparable ones of modern island arcs, then the island must have drifted or been rotated to its present position within the last 5—10 m.y.

I. Introduction

That the uniqueness of island arc volcanism is in some way a function of "lithosphere" descending into the upper mantle at the site of ocean trenches is an important corollary of the New Global Tectonics (Isacks, Oliver, and Sykes, 1968). Detailed petrochemical descriptions of island arc rocks are needed in order to assess their spatial and temporal variations and possible source regions.

This paper calls attention to the diversity of island arc rock types and delineates their chemical characteristics in one area where all are represented. It then uses this petrochemical history to comment on the tectonic development of that area.

II. Tectonic Setting

The relationship between Fiji and present components of island arcs (trench, dipping seismic plane, etc.) is unclear. However, its apparently integral association with the New Hebrides-Tonga-Kermadec-Macquarie Ridge arc system and its geological similarities to these and other circum-Pacific regions suggest that its history has been that of an island arc (Dickinson, 1967).

The Fiji Archipelago occupies part of a triangular plate in the midst of the tectonically complex southwest Pacific (Fig. 1). The plate appears to lie between two transform faults of the arc-arc type (Wilson, 1965) both connecting oppositely facing arcs. Its northern boundary is the Hunter Fracture Zone spearating the Tonga-Kermadec and New Hebridean arcs (Isacks *et al.*, 1969; Sykes *et al.*, 1969); its southern boundary is the Alpine fault zone of New Zealand separating the Tonga-Kermadec and Macquarie Ridge arcs (Karig, in prep.; Summerhayes, 1967; Banghar and Sykes, 1969).

The archipelago is situated on the northern half of a roughly N-S bathymetric high (Lau-Colville Ridge, Fig. 1) on the eastern side of the plate, parallel to but

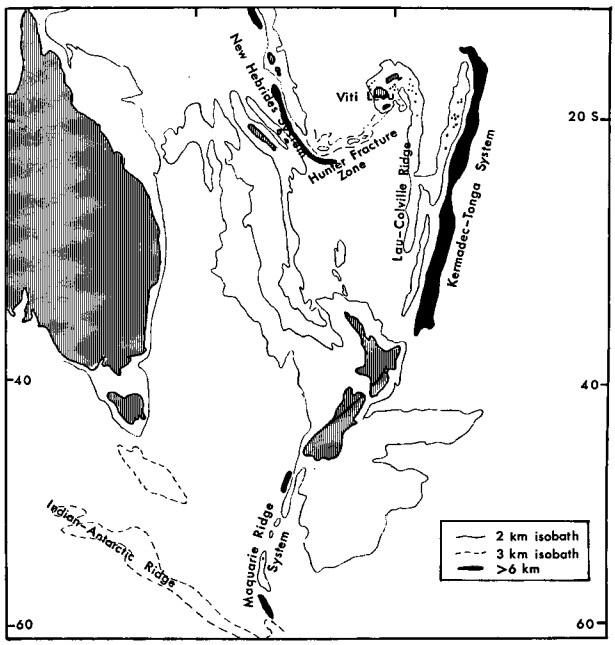

Fig. 1. Bathymetric chart of the southwest Pacific (after Cullen, 1967 and Scripts Inst. of Oceanography, 1968)

west of the Tonga-Kermadec Ridge. The high is assymetric in cross section with a sediment fan to the west but an abrupt escarpment on the eastern slope (Karig, 1970), and has a crustal thickness of about 15 km in the south (Shor, in prep.).

Its northern termination marks its intersection with the Hunter Fracture Zone. Here the Lau-Colville Ridge is sharply hooked or bent to the southwest along the strike of the New Hebrides-Tonga connection and here the largest Fiji islands, including Viti Levu, occur. The fracture zone is typified by prominent *en echelon* troughs and ridges and variable heat flow (Sclater and Menard, 1967), and considerable seismic activity (Sykes *et al.*, 1969). Earthquake focal mechanism solutions suggest that both sinistral strike-slip and thrust movements are significant (Sykes *et al.*, 1969; Isacks *et al.*, 1969). The NE-SW tectonic grain of the fracture zone is also the dominant lineament in the larger Fiji islands as seen by the structural trends of Viti Levu (Rodda, 1967; Dickinson, 1967) and the shape of Vanua Levu, Taveuni, Kandavu, and other islands.

III. Geological History

The geological history of Viti Levu has been summarized by Rodda (1967). The island was once thought part of a foundering "Melanesian continent" (Woolnough, 1903; Ladd, 1934) and does have a 25—30 km crustal thickness comparable to continental areas (Robertson, 1967; Solomon and Biehler, 1969). Igneous rocks are important in seven stratigraphic "Groups". Because relevant petrological descriptions are found only among Fiji Geological Survey publications, they are briefly summarized below.

A. Wainimala Group

Foraminifera from interbedded limestones indicate that these rocks range in age from Upper Eocene (Tertiary *b*; Cole, 1960) to Lower or Middle Miocene (Tertiary *e* and *f*; Rodda, 1967). They are largely volcanic and most are considered submarine. Greenschist facies metamorphism is widespread, a common assemblage being plagioclase-augite-chlorite ± sphene, epidote, calcite, quartz, sericite, amphibole, or zeolites. Only hornfelses at pluton contacts show significant textural modification (Crook, 1963).

There is greater apparent petrologic variety in this Group than any other, the rocks being described as spilites, keratophyres, basalts, andesites, and rhyolites (Houtz, 1960, 1963; Bartholomew, 1960; Skiba, 1964; Hirst, 1965; Band, 1968). In those described as basalt, augite and zoned plagioclase with labradorite cores are the most common phenocrysts with some reported olivine. Andesites also have calcic plagioclase and augite, lack olivine, and some have primary hornblende or hypersthene. Spilites and keratophyres have albite phenocrysts together with either augite and infrequent pseudomorphed olivine or occasional quartz but no ferromagnesian minerals (Houtz, 1960; Skiba, 1964; Band, 1968). There seems to be a continuum between "basalt" and "spilite", most rocks of basaltic texture having varying amounts of greenschist facies minerals. Fresh labradorite, calcic relicts, and albite phenocrysts are all recorded occurring with fresh or altered augite, chlorite, epidote, or sphene.

B. Tholo Plutonics

Viti Levu plutons, which range from olivine gabbro to quartz tonalite, differ from those usually described in orogenic areas by being devoid of K-feldspar. They are frequently rimmed with fine-grained diorite containing xenoliths of resorbed Wainimala Group rocks (Houtz, 1963; Hirst, 1965); the pluton interiors are most commonly medium-grained tonalite with strongly zoned plagioclase having rims of intergrown albite-oligoclase and quartz, 10—30% quartz, and $\leqq 10\%$ ferromagnesian minerals, usually hornblende and/or biotite but occasionally hypersthene (Hirst, 1965). Epidote, chlorite, calcite, and sericite are frequently found.

These plutons intrude the Wainimala Group but no others, the younger groups frequently containing plutonic detritus (Rodda, 1967). Their emplacement has been dated between 10—50 m.y. (Rodda *et al.*, 1967) but most measurements congregate at the younger figure. Dickinson (1967) and Houtz (1960) argue that Tholo plutonism and Wainimala volcanism are contemporaneous and therefore genetically related aspects of the same orogenic event.

C. Savura Volcanic Group

Volumetrically less significant than the other Groups discussed, these unconformably overlie Wainimala Group strata and range in composition from basalt to rhyolite, andesite conglomerate being most common (Rodda, 1967; Band, 1968). Greenschist facies assemblages are not reported. Their age is uncertain but assumed to be Miocene (Ibbotson, 1960).

D. Mendrausuthu Andesitic Group and Nandi Sedimentary Group

These overlie a major mid-Miocene unconformity and are Upper Miocene to Lower Pliocene (Tertiary *g*) in age (Ibbotson, 1960; Bartholomew, 1960; Rickard, 1963). In both, augite or hornblende andesite is the sole igneous rock type. The Namosi Andesite formation of the former Group is characterized by augite andesite throughout most of its history, giving way to rocks with up to 15% green or brown pleochroic hornblende higher in the sequence (Rodda, 1967; Band, 1968). Hornblende phenocrysts are common to all andesites described in the Nandi Group, though usually as less abundant than augite (Rickard, 1963; Houtz, 1963).

E. Koroimavua Andesitic Group

These rocks unconformably overlie those of the Nandi Group, yield Upper Pliocene fossils (Bartholomew, 1960), and have been dated at about 5 m.y. (McDougall, 1963). Associated monzonite plugs have comparable radiometric ages and are thought to be eruptive centers for the volcanism (McDougall, 1963; Dickinson, 1968a). Augite and strongly zoned labradorite

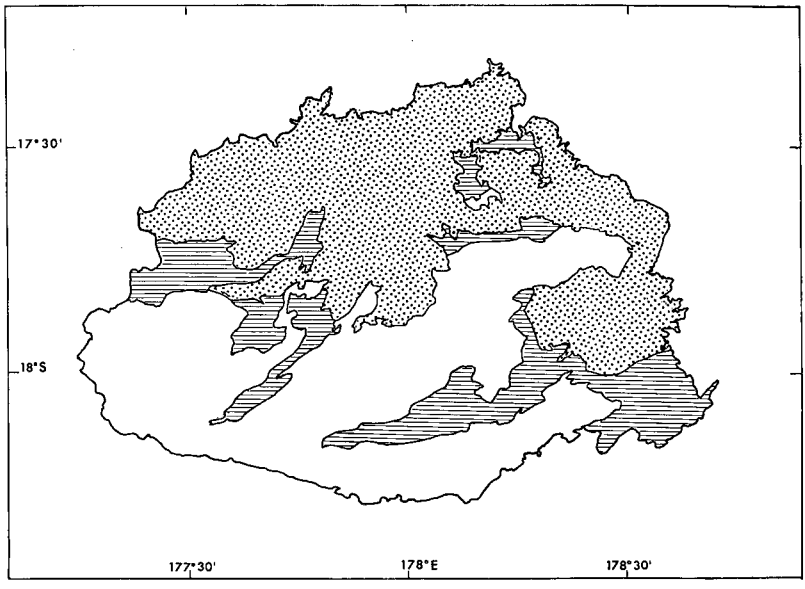

Fig. 2. Simplified geologic map of Viti Levu, Fiji. Map units correspond to rocks of "periods" defined in the text and include volcaniclastic material derived from volcanism of the respective period. The blank area indicates rocks of the first period; striped area, rocks of the second period; and dotted area, rocks of the third period. Based on Rodda and Band (1966)

plagioclase are the most abundant phenocrysts with varying amounts of hypersthene, hornblende, and biotite (Bartholomew, 1959; Rickard, 1963; Dickinson, 1968a). Leucite has also been reported (Rickard, 1963). These rocks have been described as shoshonites by Dickinson *et al.* (1968).

F. Mba Basaltic Group

Olivine-augite basalts are most typical of this Group. Plagioclase with bytownite and labradorite cores, fresh pale green augite, and usually altered olivine are the most common phenocrysts in that order of abundance; olivine and augite are sometimes absent. Trachyandesites, trachybasalts, and monzonite sills and stocks often containing groundmass sanidine are associated with the late stages of the Tavua eruptive center (Rodda, 1969, pers. comm.). Hornblende and subsidiary augite andesites occur on the Vatia Peninsula (Ibbotson, 1963) and near Rariraki volcano (Seeley, 1967).

Interbedded fossiliferous horizons consistently yield Upper Pliocene (Tertiary *h*) assemblages. Most radiometric age determinations converge between 4—5 m.y. but dates as old as 10 m.y. are recorded (Rodda *et al.*, 1967).

G. Summary

Three periods of igneous activity emerge on Viti Levu which are distinguishable in both space and time. The oldest (Wainimala, Tholo, and Savura Groups) are a melange of eruptive rocks ranging from basalt to rhyolite with common spilite and keratophyre, and plutonic rocks ranging from olivine gabbro to quartz tonalite. Most of them are characterized by an overprint of greenschist mineralogy. The youngest rocks (Koroimavua and Mba Groups) are olivine basalts and associated andesites and other rocks, occasionally containing biotite, leucite, or groundmass

sanidine. Intermediate between the two are hornblende and augite andesites (Nandi and Mendrausuthu Groups) typical of the circum-Pacific calc-alkaline suite. Spatially, the oldest rocks occupy roughly the southern half of the island, the youngest rocks the northern half, and the intermediate ones are in between (Fig. 2).

IV. Analytical Methods

Most major elements and Rb, Sr, Ba, Pb, Y and Zr were determined by the X-ray fluorescence methods of Norrish and Chappell (1967), K and Na by flame photometry as described by Cooper (1963), and FeO and loss on ignition by conventional wet chemical methods. K, U, and Th were determined by gamma-ray spectrometry as described by Heier and Rogers (1963). Measurements of Cu, Co, Ni, V, Sc, Cr, Y and Zr were made by emission spectrography as described by Ahrens and Taylor (1961). Determination of Sn, Sb, Ba, Cs, Hf, U, Th, and the rare earths utilized the spark source mass spectrographic method of Taylor (1965). The spectrographic and gamma-ray techniques provide precision, expressed as relative deviations, of at least $\pm 10\%$; the precision of other methods is better than $\pm 5\%$, expressed in the same fashion. All samples were analyzed in duplicate or triplicate. Ba, U, Th, Y, and Zr were measured by two different techniques and K by three, the valües agreeing within the levels of precision claimed. The techniques were calibrated with international standards using the values listed in Appendix 1.

The present day strontium isotope compositions were measured using the techniques described by Compston *et al.* (1965) and Arriens and Compston (1968). The standard error of seven duplicate analyses in this study was ± 0.0001. The ratios have been normalized to 0.1194 for Sr 86/88. The present day are also the initial Sr 87/86 ratios, within experimental error, even if one assumes a 40 m.y. age for the Wainimala Group samples. The 87/86 ratio for the M. I. T. Sr standard measured in this laboratory is 0.7081 (Arriens and Compston, 1968).

V. Geochemistry of the Igneous Rocks

The three periods of igneous activity distinguished in space and time also have different geochemical characteristics. Analytical data for twenty-four samples are given in Tables 1—3; sample descriptions and locations are provided in Appendix 2. Rodda (1969) has collected about fifty other published and unpublished analyses of Viti Levu rocks and although the following discussions deal specifically with the new analyses reported here, they also utilize these additional data.

A. First Period (Wainimala, Tholo and Savura Groups)

1. Volcanic Members

All of the Wainimala Group samples studied contain some greenschist facies minerals, notably chlorite and calcite. All except 68-52 contain fresh or altered zoned calcic plagioclase although only in samples 68-48 and 68-49 do total phenocrysts exceed 20 modal per cent. Despite the prevalent alteration, textural relations are maintained. None of the Savura Group samples have greenschist assemblages.

The samples are characterized by low K_2O values and a low slope on a K_2O—SiO_2 variation diagram (Fig. 3). They are iron enriched (Fig. 4a), having a ferrofemic index (Coats, 1968) between 68-80. Na_2O/K_2O ratios are variable and high (3—17, average is 6).

All samples have low concentrations of most trace elements studied, particularly the large cations. K/Rb ratios are high, averaging 1069. U and Th are low, less

Table 1a. *First period: volcanic suite*

	68-61	68-62	68-49	68-52	68-54	68-48	68-60	68-55	68-50	68-51
SiO_2	45.50	45.20	50.47	54.60	54.77	56.00	56.38	57.44	71.84	71.86
TiO_2	1.08	1.11	0.71	1.73	0.57	0.62	1.03	1.25	0.50	0.50
Al_2O_3	18.00	18.28	17.82	16.20	17.98	19.11	14.40	15.62	12.87	12.79
Fe_2O_3	7.48	8.85	3.72	4.66	4.17	2.92	8.06	3.48	1.56	1.85
FeO	3.68	2.66	5.82	4.06	3.81	4.13	2.42	5.01	2.05	1.77
MnO	0.23	0.21	0.19	0.24	0.21	0.15	0.14	0.21	0.12	0.12
MgO	5.95	5.04	4.71	3.32	3.54	2.25	3.33	3.38	0.76	0.67
CaO	9.60	9.03	7.04	3.35	8.00	8.53	5.87	6.14	3.60	2.64
Na_2O	2.69	2.78	3.93	6.71	2.63	2.95	3.16	4.20	4.68	4.44
K_2O	0.93	0.74	1.19	0.39	0.66	0.77	0.29	0.43	0.89	1.09
P_2O_5	0.29	0.30	0.06	0.58	0.18	0.10	0.12	0.44	0.12	0.11
Loss	4.32	4.67	3.81	3.54	3.14	2.21	4.24	1.99	1.79	1.83
TOTAL	99.75	98.87	99.47	99.38	99.66	99.84	99.44	99.59	99.78	99.67
F	52.0	55.0	48.0	44.0	53.0	56.0	59.0	40.0	35.0	36.0
M	30.0	26.0	25.0	18.0	25.0	18.0	20.0	21.0	8.0	7.0
A	18.0	19.0	27.0	38.0	22.0	26.0	21.0	29.0	57.0	57.0
Na_2O/K_2O	2.9	3.8	3.3	17.0	4.0	3.8	11.0	9.8	5.3	4.1
C I P W NORMS										
q	0.10	1.78		2.94	14.33	13.70	20.55	13.33	41.00	35.12
or	5.50	4.37	7.03	2.30	3.90	4.55	1.71	2.54	5.26	6.44
ab	22.76	23.52	33.26	56.78	22.26	24.96	26.74	35.54	29.36	37.57
an	34.30	35.22	27.47	12.93	35.31	26.63	24.25	22.50	12.19	11.75
ne										
di	9.05	6.09	5.78	0.22	2.66	4.19	3.24	4.36		0.57
hy	10.62	9.73	14.03	9.36	10.54	7.97	6.79	10.95	3.77	2.51
ol			1.23							
he	0.94	4.68					4.42			
mt	9.48	6.04	5.39	6.76	6.05	4.23	5.27	5.05	2.26	2.68
il	2.05	2.11	1.35	3.29	1.08	1.18	1.96	2.37	0.95	0.95
ap	0.63	0.66	0.13	1.27	0.39	0.22	0.26	0.96	0.26	0.24
Cs	—	—	—	< 0.1	< 0.1	< 0.1	< 0.1	—	0.3	—
Rb	7.9	4.7	5.8	3.0	9.2	7.0	2.1	4.0	6.8	8.9
Ba	161	179	124	50	184	109	70	50	218	279
Sr	591	492	109	130	360	168	128	219	116	118
Pb	3.0	3.0	1.3	11	3.2	3.9	1.9	1.2	1.2	1.0
K/Rb	982	1,302	1,700	1,077	600	908	1,138	890	1,082	1,016
Rb/Sr	0.01	0.01	0.05	0.02	0.03	0.04	0.02	0.01	0.06	0.08
Ba/Rb	20	38	21	17	20	16	32	12	32	31
Th	0.60	0.75	n.d.	0.22	< 0.10	n.d.	0.17	0.31	< 0.10	< 0.10
U	0.23	0.16	n.d.	0.17	< 0.10	< 0.10	< 0.10	0.34	0.17	0.22
Zr	62	68	100	103	54	55	68	74	71	72
Hf	—	—	—	0.82	1.1	1.6	1.5	—	4.5	—
Sn	—	—	—	0.98	0.91	0.78	1.3	—	2.9	—
Th/U	2.6	4.7	—	1.3	1.0	—	—	0.9	0.3	0.3
Zr/Hf	—	—	—	126	49	34	—	—	16	—

Table 1a (continued)

	68-61	68-62	68-49	68-52	68-54	68-48	68-60	68-55	68-50	68-51
La	—	—	—	5.9	3.9	2.3	2.4	—	2.7	—
Ce	—	—	—	18	10	4.6	7.2	—	7.2	—
Pr	—	—	—	3.0	1.8	1.1	1.4	—	2.1	—
Nd	—	—	—	17	9.1	7.0	8.4	—	14	—
Sm	—	—	—	3.9	2.8	3.5	3.5	—	6.5	—
Eu	—	—	—	1.9	1.1	1.1	1.2	—	2.2	—
Gd	—	—	—	2.4	1.7	2.0	1.9	—	4.5	—
Tb	—	—	—	0.49	0.35	0.51	0.52	—	1.1	—
Dy	—	—	—	3.3	2.3	3.4	3.5	—	7.3	—
Ho	—	—	—	0.76	0.55	0.90	0.89	—	1.9	—
Er	—	—	—	1.5	1.3	2.1	1.8	—	4.9	—
Tm	—	—	—	0.18	0.17	0.20	0.23	—	0.64	—
Yb	—	—	—	1.7	1.6	2.6	2.5	—	6.0	—
Y	22	23	21	41	18	28	34	30	43	44
Σ	—	—	—	101	54.7	59.3	65.4	—	104	—
La/Yb	—	—	—	3.5	2.4	0.88	0.96	—	0.45	—
Cu	87	78	75	2	23	31	77	23	n.d.	2
Ni	19	11	6	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Co	38	32	31	8	15	18	27	16	n.d.	3
V	290	250	260	74	91	150	190	150	9	9
Sc	37	34	37	23	13	28	35	26	16	18
Cr	31	7	15	n.d.	11	n.d.	2	2	n.d.	2
Sb	—	—	—	0.19	0.33	0.24	n.d.	—	n.d.	—
Ni/Co	0.5	0.3	0.2	—	—	—	—	—	—	—
$Sr^{87/86}$	0.7040	0.7038	0.7053	0.7034	0.7040	0.7040	0.7050	0.7037	0.7049	0.7043

— = not determined; n.d. = not detected.

than 0.3 ppm in all but two Savura Group samples, Th/U ratios being near unity. Ferromagnesian elements are generally less abundant than in other Fiji rocks, Ni being undetected in most samples.

The rare earth abundance pattern is subparallel to that of chondrites (Fig. 5), being somewhat depleted in the larger elements (La-Pr). Concentrations are low, showing about a ten-fold enrichment relative to chondrites. There is a scatter in light rare earth element concentrations, although each sample has its greatest enrichment relative to chondrites in the Nd-Eu range. All but one have quite uniform abundances of the heavier elements (Tb—Yb) and the La/Yb ratio is 0.5—3.5. The exception (68-50) is a "rhyolite" ($SiO_2 = 72\%$), devoid of ferromagnesian minerals.

The average initial Sr 87/86 ratio is 0.7042, although the range (0.7034—0.7053) exceeds experimental error, and is greater than for other periods.

2. Plutonic Members

Chlorite, epidote, calcite, sphene, and sericite make up 10—20 modal per cent of all samples investigated. The three dikes have gabbroic mineralogies although 68-58 also contains free quartz; the tonalite has almost 40% modal quartz but no K-feldspar.

Table 1b. *First period: Plutonic suite*

	68-57	68-56	68-58	68-59
SiO_2	49.55	50.06	50.88	75.32
TiO_2	1.31	1.41	1.22	0.23
Al_2O_3	16.40	16.45	15.07	12.58
Fe_2O_3	2.86	3.47	3.79	1.40
FeO	6.00	5.74	7.50	0.30
MnO	0.21	0.22	0.19	0.04
MgO	6.95	6.81	4.75	0.43
CaO	9.79	9.67	8.51	2.01
Na_2O	2.76	2.86	3.28	5.59
K_2O	0.21	0.19	0.35	0.24
P_2O_5	0.27	0.28	0.10	0.04
Loss	2.82	2.80	3.55	0.84
TOTAL	99.13	99.96	99.19	99.02
F	46	47	57	20
M	38	36	25	5
A	16	17	18	75
Na_2O/K_2O	13	15	9.4	23
q	2.19	3.12	4.82	37.30
or	1.24	1.12	2.07	1.42
ab	23.36	24.20	27.76	47.30
an	31.74	31.49	25.36	8.53
ne				
di	12.26	11.90	13.34	0.94
hy	18.29	17.02	14.27	0.63
ol				
he				1.10
mt	4.15	5.03	5.50	0.43
il	2.49	2.68	2.32	0.44
ap	0.59	0.61	0.22	0.09
Cs	0.16	—	—	< 0.1
Rb	1.6	1.6	4.9	1.9
Ba	86	89	44	42
Sr	252	252	164	123
Pb	1.2	1.7	3.4	0.3

	68-57	68-56	68-58	68-59
K/Rb	1,075	975	585	1,068
Rb/Sr	0.01	0.01	0.03	0.02
Ba/Rb	54	56	9	22
Th	0.19	0.18	< 0.1	2.7
U	0.18	0.12	0.15	1.1
Zr	121	128	72	122
Hf	1.7	—	—	3.8
Sn	1.9	—	—	2.1
Th/U	1.0	1.3	< 1.0	2.4
Zr/Hf	71	—	—	32
La	9.5	—	—	11
Ce	18	—	—	17
Pr	3.2	—	—	2.7
Nd	16	—	—	12
Sm	5.7	—	—	3.9
Eu	1.8	—	—	0.92
Gd	3.0	—	—	2.7
Tb	0.68	—	—	0.45
Dy	4.6	—	—	2.7
Ho	1.1	—	—	0.62
Er	2.6	—	—	1.8
Tm	0.28	—	—	0.25
Yb	2.9	—	—	2.3
Y	28	30	26	17
Σ	97	—	—	75
La/Yb	3.3	—	—	4.8
Cu	39	48	140	< 1
Ni	82	72	16	3
Co	33	30	32	4
V	210	230	260	26
Sc	26	27	27	4
Cr	170	170	12	2
Sb	nd	—	—	0.36
Ni/Co	2.5	2.4	0.5	0.7
Sr 87/86	0.7033	0.7047	0.7034	0.7044

Each of these samples is generally richer in soda than those of comparable silica content in the volcanic suite, but they are indistinguishable from them on K_2O—SiO_2 and FMA diagrams (Figs. 3 and 4a). Several trace elements also show similarities to the volcanic rocks: Rb, Sr, Ba, Pb, and Cs are comparably low, K/Rb ratios average 925, and Rb/Sr ratios are 0.01—0.03. The three basic dikes have U and Th concentrations and Th/U ratios similar to those of the volcanic suite although the tonalite does not.

There are other significant differences between the volcanic and plutonic rocks. Ferromagnesian elements are generally more abundant in the basic dikes but comparable in the tonalite. Zr is uniformly high, presumably due to a stable

Table 2. *Second period*

	66	X88[a]	X96[a]
SiO_2	59.39	59.6	57.4
TiO_2	0.68	0.70	0.65
Al_2O_3	16.73	17.1	17.7
Fe_2O_3	3.66	2.95	3.20
FeO	2.61	2.90	3.30
MnO	0.13	0.13	0.15
MgO	3.08	3.15	3.45
CaO	7.12	7.25	7.40
Na_2O	3.97	3.75	3.40
K_2O	1.27	1.25	1.25
P_2O_5	0.20	—	—
Loss	0.55	0.67	1.60
TOTAL	99.39	99.45	99.50
F	41	40	43
M	22	23	24
A	37	37	33
Na_2/K_2O	3.1	3.0	2.7
q	13.86	13.99	12.23
or	7.51	7.39	7.39
ab	33.59	31.73	28.77
an	24.08	26.14	29.34
ne			
di	7.93	7.83	5.89
hy	4.83	6.09	8.40
ol			
he			
mt	5.31	4.28	4.64
il	1.29	1.33	1.23
ap	0.44		
Cs	—	0.08	0.47
Rb	16	22	22
Ba	462	330	270
Sr	481	445	540
Pb	3.9	3.0	2.2

	66	X88	X96
K/Rb	657	473	474
Rb/Sr	0.03	0.05	0.04
Ba/Rb	29	15	12
Th	2.0	2.4	1.4
U	0.71	0.49	0.29
Zr	111	115	95
Hf	—	3	1.9
Sn	—	0.36	0.29
Th/U	2.9	4.9	4.8
Zr/Hf	—	38	50
La	—	9.7	9.1
Ce	—	17	17
Pr	—	2.6	3.0
Nd	—	8.8	11
Sm	—	2.1	2.1
Eu	—	0.63	0.70
Gd	—	1.8	1.7
Tb	—	0.32	0.27
Dy	—	—	—
Ho	—	0.43	0.36
Er	—	1.6	1.2
Tm	—	0.25	0.17
Yb	—	1.1	0.88
Y	17	20	18
Σ	—	66	65
La/Yb	—	8.8	10.3
Cu	43	30	—
Ni	11	8	—
Co	16	16	—
V	170	160	—
Sc	14	18	—
Cr	31	36	—
Sb	—	—	—
Ni/Co	0.7	0.5	—
Sr 87/86	0.7041	—	—

[a] Taylor *et al.* (1969).

zircon phase. The rare earth abundance patterns of the tonalite and one dike are quite different than those of the volcanics, being relatively more enriched in the lighter elements (Fig. 6).

3. Influence of Metamorphism

Because all of the Wainimala Group samples studied are spilitic, i.e., "rocks of basaltic texture in which greenschist facies mineralogy is completely or almost completely developed" (Cann, 1969), the question arises whether these chemical characteristics are primary or acquired.

Table 3. *Third period*

	68-72	68-74	68-71	68-75	68-63	68-64	68-65	68-76	68-73
SiO_2	48.37	48.80	49.43	49.67	50.50	50.80	50.79	51.58	52.65
TiO_2	0.58	0.18	1.00	0.73	0.54	0.62	0.62	0.53	0.72
Al_2O_3	12.45	11.94	15.82	17.32	17.04	19.89	19.89	20.05	17.36
Fe_2O_3	4.89	4.41	5.92	5.15	4.65	3.61	3.53	4.07	7.00
FeO	5.33	4.79	4.50	3.73	2.42	3.65	3.79	2.16	2.02
MnO	0.20	0.18	0.20	0.14	0.17	0.15	0.14	0.15	0.17
MgO	10.53	10.39	5.77	5.90	3.36	3.70	3.61	2.05	3.29
CaO	11.65	11.44	10.08	10.51	8.98	9.07	8.95	7.73	7.70
Na_2O	2.59	2.81	2.89	2.45	4.96	2.90	3.03	3.32	3.43
K_2O	1.75	1.52	2.71	2.12	4.00	3.28	3.46	3.99	3.90
P_2O_5	0.34	0.40	0.38	0.25	0.44	0.44	0.45	0.50	0.56
Loss	1.83	3.02	1.41	2.03	3.33	1.80	1.40	2.95	1.96
TOTAL	100.51	99.86	100.11	100.00	100.39	99.91	99.66	99.08	100.76
F	39	37	46	44	35	41	41	38	44
M	43	44	27	31	18	22	21	14	17
A	8	9	27	25	47	37	38	48	39
Na_2O/K_2O	1.5	1.8	1.1	1.1	1.2	0.9	0.9	0.8	0.9
q				0.33				0.24	0.91
or	10.34	8.98	16.02	12.53	23.64	19.38	20.45	23.58	23.05
ab	16.79	20.23	23.57	20.73	19.22	24.54	24.91	28.09	29.02
an	17.18	15.48	22.19	30.00	12.42	31.57	30.45	28.02	20.45
ne	2.78	1.92	0.48		12.33		0.39		
di	30.63	30.98	20.20	16.27	20.73	8.68	9.08	5.75	11.25
hy				8.70		2.81		2.45	2.98
ol	12.03	11.68	4.93			3.75	5.70		
he									3.57
mt	7.09	6.39	8.58	7.47	6.74	5.28	5.12	5.90	4.98
il	1.10	0.34	1.90	1.39	1.03	1.18	1.18	1.01	1.37
ap	0.74	0.83	0.83	0.55	0.96	0.96	0.98	1.09	1.22
Cs	—	0.90	—	—	—	0.79	—	—	—
Rb	60	22	30	37	71	56	64	187	75
Ba	684	620	456	500	677	677	689	992	684
Sr	972	773	1,281	645	1,552	1,228	1,235	2,010	1,045
Pb	8.8	7.5	7.0	4.9	11	7.9	8.2	13	10
K/Rb	243	575	442	476	468	485	449	177	431
Rb/Sr	0.06	0.03	0.02	0.06	0.05	0.05	0.05	0.09	0.07
Ba/Rb	11	35	8	13	9	12	11	5	9
Th	0.79	1.23	1.02	0.52	2.25	1.30	0.85	1.59	1.12
U	0.36	0.41	0.49	0.18	1.23	0.56	0.54	0.68	0.48
Zr	34	54	64	26	67	46	46	57	71
Hf	—	0.8	—	—	—	1.1	—	—	—
Sn	—	1.0	—	—	—	1.6	—	—	—
Th/U	2.2	3.0	2.1	2.9	1.8	2.3	1.6	2.3	2.3
Zr/Hf	—	66	—	—	—	42	—	—	—

Table 3 (continued)

	68-72	68-74	68-71	68-75	68-63	68-64	68-65	68-76	68-73
La	—	14	—	—	—	8.5	—	—	—
Ce	—	29	—	—	—	14	—	—	—
Pr	—	3.1	—	—	—	2.2	—	—	—
Nd	—	15	—	—	—	11	—	—	—
Sm	—	3.2	—	—	—	4.0	—	—	—
Eu	—	1.1	—	—	—	1.2	—	—	—
Gd	—	1.9	—	—	—	2.4	—	—	—
Tb	—	0.36	—	—	—	0.43	—	—	—
Dy	—	2.0	—	—	—	3.0	—	—	—
Ho	—	0.33	—	—	—	0.58	—	—	—
Er	—	0.80	—	—	—	1.4	—	—	—
Tm	—	0.11	—	—	—	0.19	—	—	—
Yb	—	1.0	—	—	—	1.5	—	—	—
Y	14	14	23	16	13	12	13	16	26
Σ	—	85.9	—	—	—	62.4	—	—	—
La/Yb	—	14.0	—	—	—	5.7	—	—	—
Cu	120	91	100	79	180	200	260	150	150
Ni	92	110	17	25	6	13	13	4	7
Co	47	41	32	31	20	21	23	17	23
V	670	240	350	330	330	220	260	180	200
Sc	35	34	29	34	15	14	16	8	18
Cr	360	220	9	72	3	40	78	2	9
Sb	—	n.d.	—	—	—	0.30	—	—	—
Ni/Co	2.0	2.7	0.5	0.8	0.3	0.6	0.6	0.2	0.3
Sr 87/86	0.7040	0.7042	0.7043	0.7042	0.7043	0.7037	0.7040	0.7041	0.7039

Many of these characteristics are similar to those of abyssal tholeiites. The development of spilites from these basalts has been studied recently with the following results: 1. the rocks maintain their subalkaline and iron enriched nature (Cann, 1969; Melson *et al.*, 1969; Melson and van Andel, 1966); 2. ferromagnesian trace elements change little (Cann, 1969; Melson *et al.*, 1969); 3. rare earth abundances and distribution patterns are not affected (Frey *et al.*, 1969; Graham and Nicholls, 1969); and 4. the K/Rb ratio may be increased (Hart, 1969).

Most of these writers argue that abyssal spilite represents oceanic tholeiite altered by burial metamorphism (Melson and van Andel, 1966; Cann, 1969). Analogous arguments have also been advanced for other areas (Vallance, 1960, 1965; Smith, 1968). Fijian spilites are thought to have originated similarly (Crook, 1963) because of the common mineral assemblages in both volcanics and sediments and the maintenance of textual integrity in the volcanics.

For these reasons, the important chemical characteristics of this suite are thought to be essentially primary features, perhaps slightly modified but not sufficiently as to discredit the conclusions. This view is strengthened by the chemical similarities of the Savura Groups rocks in which no such metamorphism is recognized.

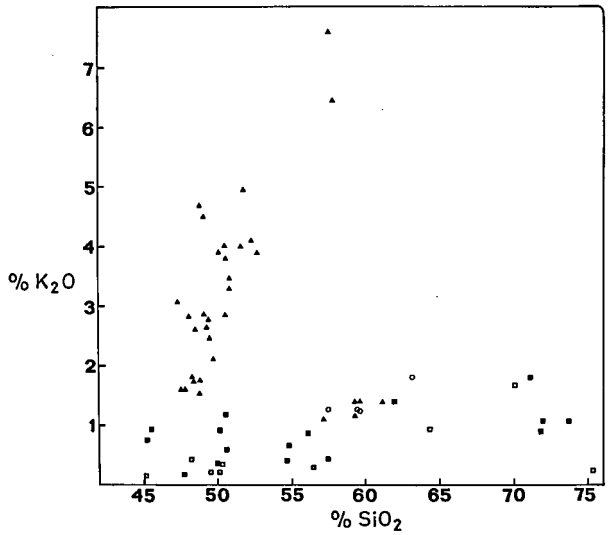

Fig. 3. K_2O—SiO_2 variation diagram for Viti Levu rocks. All available data are included: this paper, Rodda (1969), Ibbotson (1963, 1967), Seeley (1967). Solid squares indicate first period volcanic samples; open squares, first period plutonic samples; open circles, second period; solid triangles, third period; open triangles, hornblende andesite samples from Vatia Peninsula and Rakiraki volcano

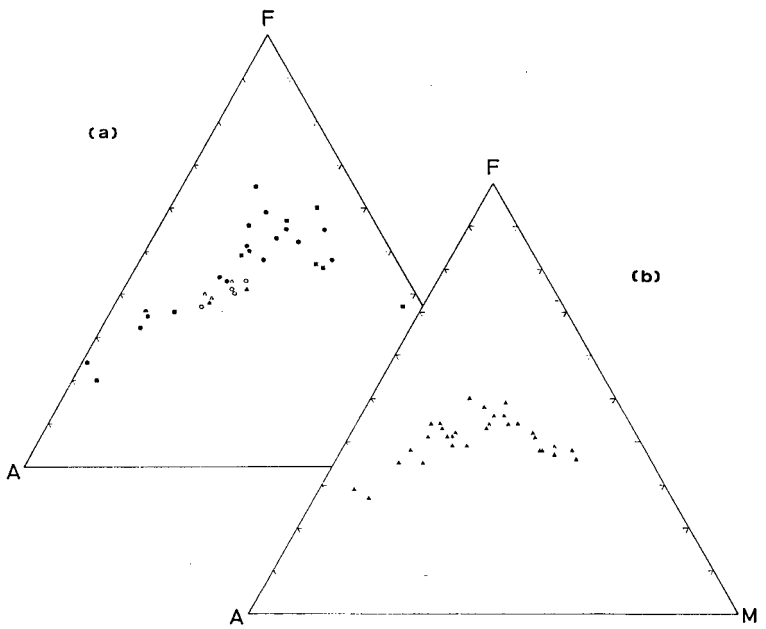

Fig. 4a and b. FMA diagram for Viti Levu rocks. F is total iron as FeO. Data as in Fig. 3. (a) first and second period samples plus those from Vatia Peninsula and Rakiraki volcano. Solid circles indicate first period volcanic samples; solid squares, first period plutonic samples; open circles, second period; open triangles, Vatia and Rakiraki samples. (b) third period samples

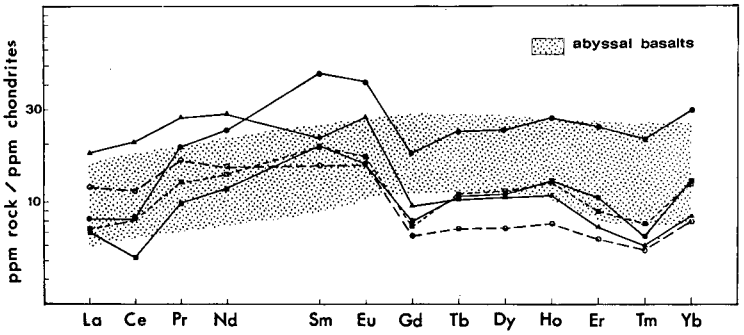

Fig. 5. First period volcanic rock rare earth abundance patterns. Abundances are normalized relative to their respective concentrations in the composite of 9 chondrites (Frey *et al.*, 1968). The field of abyssal basalts is reproduced from Gast (1968) for comparison. Open triangles, 68-52; open circles, 68-54; solid circles, 68-50; open squares, 68-60; solid squares, 68-48

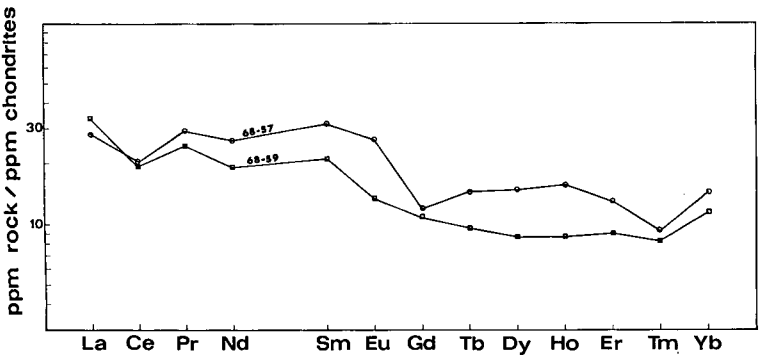

Fig. 6. First period plutonic rock rare earth abundance patterns

B. Second Period (Mendrausuthu and Nandi Groups)

Both of the Fiji andesites described by Taylor *et al.* (1969) as "typical circum-Pacific andesites" are from this suite. One additional analysis is added here and it is similar to their sample X-88 in mineralogy and chemistry (Table 2).

In contrast to rocks of the first period, these have higher K_2O values and a steeper slope on a K_2O—SiO_2 variation diagram (Fig. 3). No iron enrichment is seen (Fig. 4a). Na_2O/K_2O ratios are 2.7—3.1.

The large cations, Th, and U are 2—4 times more abundant than in the earlier period. K/Rb ratios range between 450—550 and Th/U ratios between 3—5. Ni and Cr concentrations are higher, being about 10 and 33 ppm respectively. The rare earth concentrations (Fig. 7) show some relative enrichment of the lighter elements and abundance patterns similar to those of Hawaiian tholeiites (Hubbard, 1969). La/Yb ratios are 9—10. One initial Sr 87/86 ratio is 0.7041.

C. Third Period (Koroimavua and Mba Groups)

Two new analyses of Korimavua and seven of Mba Group samples are reported (Table 3). Labradorite and augite phenocrysts are common to all, ranging from

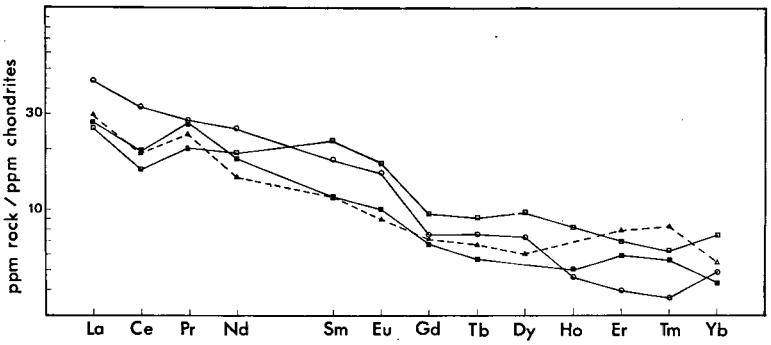

Fig. 7. Second and third period rare earth abundance patterns. Data for X 96 and X 88 are from Taylor *et al.* (1969). Open circles 68-74; solid squares, X 96; open triangles, X 88; open squares, 68-64

<1—60% and <1—40% respectively. Olivine, altered to pseudomorphed, is found in several samples; biotite phenocrysts comprise about 5% of 68-63.

The striking geochemical features of these rocks are their potassium enrichment and steep slope on a K_2O—SiO_2 variation diagram (Fig. 3). Na_2O/K_2O ratios average about 1.0 as they also do in analyses of six other Koroimavua (Dickinson *et al.*, 1968) and twelve Mba Group samples (Ibbotson, 1967). Slight iron enrichment is encountered (Fig. 4b).

As expected, large cations are most abundant in these rocks. K/Rb ratios range between 175—575, averaging 416; Ba/Rb ratios are lower (12.5) and Rb/Sr slightly higher (0.05) than in other suites. Th and U are generally enriched in 68-63 but in others average about 1.0 and 0.5 ppm respectively. The Ni/Co ratio is variable, being high in olivine-bearing samples but less than 1.0 in others. Rare earth concentrations and abundance patterns are quite similar to those of the calc-alkaline suite (Fig. 7).

The average initial Sr 87/86 ratio is 0.7041 and the range between 0.7037—0.7045.

VI. Discussion

Island arcs are commonly considered to be calc-alkaline provinces where andesite predominates with subordinate amounts of basalt, dacite, and rhyolite. This generalization is supported by the broad mineralogical similarity of rocks through time in such areas, but it may, nonetheless, obscure important time-dependent variations in arc evolution. It assumes that, except for lateral variations perpendicular to the arc axis (Kuno, 1966; Dickinson, 1968b) and the effects of differentiation, island arc magmas have remained essentially isochemical through time.

This is clearly not true in Viti Levu. Rocks of the second period correspond closely to the "typical calcalkaline series" (e.g., Nockolds, 1954): $SiO_2 = 57$—63%, no iron enrichment, moderate K_2O and moderate slope on a K_2O—SiO_2 variation diagram, Na_2O/K_2O ratios of 2—3, and trace element concentrations similar to

those described by Taylor and White (1966) and Taylor (1968) for "average andesites".

Although comparable in mineralogy, rocks of the first and third periods are quite different chemically. They have in common the low TiO_2 and high Al_2O_3 contents of orogenic areas and initial Sr 87/86 ratios of about 0.704, but differ in K_2O—SiO_2 relationships, Na_2O/K_2O ratios, degree of iron enrichment, and trace elements.

Volcanic rocks of the first period are a submarine assemblage associated with deep water sediments, spilites, and keratophyres. They display a wide range of SiO_2 contents, have lower K_2O concentrations, a flatter slope on a K_2O—SiO_2 variation diagram, and greater iron enrichment in middle fractionation stages. (Na_2O/K_2O ratios are higher but this may not be significant in such spilitized samples.) Therefore, they are called "island arc tholeiites" (for definition and discussion see Jakes and Gill, in prep.). Their trace element chemistry is also different from calcalkaline rocks and is, instead, analogous to that of abyssal tholeiites (Engel *et al.*, 1965; Tatsumoto *et al.*, 1966; Hubbard, 1969): high K/Rb ratios, low Th and U contents and Th/U ratios around 1, Sr generally less than 200 ppm, and rare earth abundance patterns subparallel to those of chondrites with La/Yb ratios of about 1. In contrast, however, Fijian "island arc tholeiites" have much less Ni and Cr, lower Ni/Co ratios (< 1), more Ba, and higher initial Sr 87/86 ratios than their ocean ridge analogues.

Relative to the second period calcalkaline series, rocks of the third period have considerably higher concentrations of Na, K, and the trace elements which substitute for them. They also have a steeper slope on a K_2O—SiO_2 variation diagram, Na_2O/K_2O ratios around 1, a Peacock alkali-lime index of 51—52 rather than > 60, and very slight iron enrichment. However, they also differ from alkali basalts of the intra-oceanic region (MacDonald and Katsura, 1964; Engel *et al.*, 1965; McBirney and Aoki, 1968; Hubbard, 1969) by having as much K_2O as Na_2O at all silica contents with corresponding phenocrysts of biotite and (rarely) leucite, much less TiO_2 and Zr, less iron enrichment, and generally more calcium. The rare earth abundance patterns are much more similar to those of calcalkaline rocks or Hawaiian tholeiites than alkali basalts.

For these reasons rocks of the third period are considered to be "island arc alkali basalts" or, preferably, shoshonites (Dickinson *et al.*, 1968). They differ from other shoshonites (Joplin, 1968; Jakes and White, 1970) by being slightly more enriched in total iron, MgO, and CaO, by having a lower alkali-lime index, normative nepheline in fifteen of twenty-five available analyses, and a steep rather than flat slope on a K_2O—SiO_2 variation diagram.

The calcalkaline rocks which one associates with island arcs are, therefore, only one of three quite different rock types on Viti Levu and volumetrically quite possibly the least important. Some similar rocks do occur amidst the Mba Group shoshonites at Vatia Peninsula and Rakiraki volcano which may indicate that their spatial and temporal distribution is greater than implied above. It is interesting that those on Vatia Peninsula are older than most Mba Group samples (Rodda *et al.*, 1967) and of comparable age to that inferred for rocks of the second period.

VII. Comparison with Other Island Arcs

Rocks here described as "island arc tholeiites" are the earliest and most abundant rocks in Fiji. (Preliminary information (Rickard, 1966; Rodda, 1969; Gill, unpublished data) suggests that much of Vanua Levu, Lau, the Mamanuthu, and part of the southern Yasawa islands are characterized by similar material.) They are clearly different from later calcalkaline volcanics.

Examination of other island arc regions suggests that Fiji is not atypical. Table 4 presents a compilation of diagnostic information from many parts of the non-continental Pacific perimeter which is drawn upon in the following discussion.

Table 4. *Selected geochemical characteristics of some island arcs*

	Average ferro-femic index[a]	Per cent K_2O at $SiO_2 = 60\%$	Δ K_2O per 5% SiO_2 (%)	Na_2O/K_2O
Fiji, 1st Period[b]	74	0.6	0.2	6.0
Fiji, 2nd Period[b]	64	1.5	0.5	3.0
Tonga[c]	74	0.6	0.2	4.3
New Hebrides:				
a) Central Islands[d]	70	2.0	0.5	2.0
b) Malekula[e]	70	1.5	0.5	4.0
Bougainville, Solomon Islands[f]	68	1.8	0.2	1.9
New Guinea-New Britain:				
a) Coastal Provinces[g]	76	1.5	0.5	2.6
b) East Papua[h]	50	2.2	0.6	2.0
Mariana Islands[i]	75	0.8	0.2	4.3
Izu Islands and Peninsula[j]	78	0.7	0.2	5.3
Kurile Islands[k]				
a) Southern Group	75	0.8	0.2	4.2
b) Central Group	72	1.1	0.3	3.5
c) Northern Group	70	2.1	0.5	1.8
Aleutian Islands[l]	65	1.5	0.5	2.4
St. Kitts, Lesser Antilles[m]	73	0.7	0.2	7.1
South Sandwich Islands[n]	76	0.8(?)	0.2 (?)	4—7

[a] See Coats (1968) for definition.
[b] This paper; Rodda (1969).
[c] Richard (1962).
[d] Warden (1967)
[e] Mitchell (1966).
[f] Blake (1968); Taylor *et al.* (1969).
[g] Morgan (1966); Lowder and Carmichael (1970).
[h] Jakes and White (1970).
[i] Schmidt (1957); Stark (1964) (excluding Alutom fm.).
[j] Kuno (1950, 1962); Isshiki (1963).
[k] Gorshkov (1958, 1966).
[l] Average values for Adak, Achmitka, Andreanof, Buldir, Delarof, Kista, Little Sitkin, Segula, and Unalaska Islands. Ferrofemic indices range from 54—68. Semisopochnoi and Umnak Islands have approximately 0.5% more K_2O at a given SiO_2 content. Data from USGS Bulletin 1028 A-S, Coats (1952), and Byers (1961).
[m] Baker (1968a).
[n] Baker (1968b); Katsui (1969).

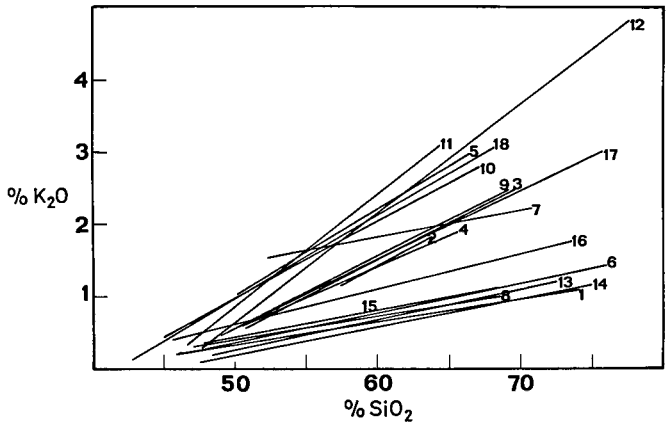

Fig. 8. Estimated K_2O—SiO_2 variation diagram for regions summarized in Table 4. Based on Dickinson (1968). Note both the continuum of slopes but also the three rather distinct populations of approximately 0.2%, 0.5%, and 0.7% K_2O per 5% SiO_2 respectively. Data as follows: *1* First period, Fiji; *2* Second period, Fiji; *3* New Britain (Lowder and Carmichael, 1970); *4* North coast New Guinea (Morgan, 1966); *5* East Papua, New Guinea; *6* Mariana Islands; *7* Bougainville, Solomon Islands; *8* St. Kitts, Lesser Antilles; *9* Malekula, New Hebrides; *10* Central Islands, New Hebrides; *11* Semispochnoi, Aleutian Islands (U.S.G.S. Bull. 1028-0); *12* Umnak, Aleutian Islands (Byers, 1961); *13* Izu Islands and Peninsula; *14* Tonga; *15* S. Kuriles; *16* C. Kuriles; *17* Aleutian Islands; *18* N. Kuriles. References not listed are found in Table 4

For comparison, the "ferro-femic index" of Nockolds' (1954) average calcalkaline series is 42; of the Cascades, 54—70; of the Thingmuli trend, 80; and of the Skaergaard, 95 (see Coats, 1968). Fig. 8, adapted from Dickinson (1968b) gives estimated K_2O—SiO_2 curves for the areas summarized in Table 4.

"Island arc tholeiites" are characteristic of Tonga, the Mariana islands, the Izu islands and peninsula, the southern Kuriles, and the Scotia arc. They have in common average ferro-femic indices between 74—78, low K_2O and K_2O—SiO_2 slopes, and high Na_2O/K_2O ratios. They are similar to Viti Levu tholeiites in some aspects of trace element chemistry. Three Saipan samples reported by Taylor *et al.* (1969) and several Izu Peninsula samples (Mazuda, 1966, 1968) all have rare earth abundance patterns subparallel to those of chondrites and La/Yb ratios of 1—2.5. (Samples of the hypersthene rock series (Taylor and White, 1966) have typical calcalkaline rare earth contents and abundance patterns.) Th and U concentrations are variable but <0.1 ppm in one Saipan andesite and 0.1—0.3 ppm in most Izu analyses (Tatsumoto, 1966). Cs, Ba, and Sr are all low in the Saipan samples. K/Rb ratios of 500—870 in Saipan and 450 in the Scotia arc (Baker, 1968b) are lower than those of Viti Levu.

In contrast, calcalkaline suites are characterized by varying iron enrichment (ferro-femic indices between 50—70), moderate K_2O—SiO_2 slopes, Na_2O/K_2O ratios of 2—3, and trace element abundances as summarized by Taylor and White (1966) and Taylor (1968). These are found in the New Hebrides, the Solomon islands, eastern Papua in New Guinea, the northern Kuriles, and the Aleutians.

Thus all island arcs are not necessarily or even principally calcalkaline provinces. Instead, the calcalkaline suite may represent but one stage in arc evolution (Jakes and White, 1970).

Although the rocks of the Lesser Antilles are generally considered calcalkaline, they are characterized by "tholeiitic" K_2O—SiO_2 variations and more iron enrichment than other calcalkaline provinces (Nockolds and Allen, 1953; Baker, 1968a; Hatherton and Dickinson, 1969). Relative to the "average andesite" of Taylor (1968), St. Kitts samples of comparable SiO_2 have less Ni and Cr as do Viti Levu tholeiites.

Rocks of the coastal regions of the New Guinea and New Britain are considered tholeiitic by Morgan (1966) and Jakes and White (1970) due to their relative iron enrichment. They also have several trace element characteristics of "island arc tholeiites": K/Rb ratios around 1000 (Jakes and White, in prep.), U and Th generally less than 0.5 ppm with Th/U ratios around 1, and low Ni and Cr contents (Ewart and Gill, unpublished data). In contrast, however, they have more K_2O and associated trace elements and a steeper K_2O—SiO_2 slope. Rocks of the Central Kuriles are intermediate between tholeiitic and calcalkaline trends in both iron enrichment and K_2O—SiO_2 variations.

In all island arc regions where shoshonites are known (Indonesia, New Guinea, Fiji, Kamchatka), they represent the last volcanic event and are furthest from the trench. In this respect they are similar to the alkali basalts of Japan and the Aleutians, but differ from them principally in Na/K relationships and corresponding mineralogy.

Of all the areas examined, however, only in Viti Levu is there a clear secular evolution from tholeiitic to calcalkaline to shoshonitic volcanism. Donnelly *et al.* (in prep.) describe a transition from submarine spilites and keratophyres to subaerial calcalkaline rocks in the Lesser Antilles. The earliest rocks are characterized by low Th and U contents (<0.5 ppm) and Th/U ratios around 1, whereas all are higher in samples of the later period. A similar change from submarine to subaerial assemblages accompanied by a decrease in iron enrichment and increase in potassium is reported from the western Aleutians (Wilcox, 1959). Baker (1968b) and Jakes and White (1970) also suggest that early arc development is tholeiitic in character, giving way to subsequent calcalkaline activity.

VIII. Conclusions

"Island arc tholeiites" are the earliest and most voluminous rocks of Fiji. They are unconformably overlain by volcanics typical of the circum-Pacific calcalkaline suite as these are in turn by shoshonites. Each rock type has distinctive features and there are marked geochemical discontinuities between the three. The latter two may overlap in time. This zonation is spatial as well as temporal, generally following the pattern of Kuno (1966) from tholeiite in the south to shoshonite in the north.

Clearly, the history of Viti Levu is that of an oceanic island arc in which magmas evolved relatively free from influences of continental crust.

The uniqueness of island arc volcanism must be related in some way to its tectonic association with the "lithosphere" thought to be thrust beneath the arcs

along Benioff zones. Green and Ringwood (1968) and Taylor *et al.* (1969), among others, have suggested that magmas closely resembling those of the second period of Viti Levu volcanism represent partial melts of down-going oceanic crust. Jakes and White (in prep.) have extended this model to account for "island arc tholeiites" and shoshonites as well, but the latter's frequent occurrence in aseismic areas (Joplin, 1968) makes its relationship to down-going crust unclear. The secular and lateral differences in magmas generated in accord with this model would be functions of the depth and degree of partial melting, the phases involved, the amount of subsequent "wall rock reaction", and the depth, P_{H_2O}, and surroundings of high-level chambers where eventual fractionation takes place.

Alternatively, the geochemical similarities between "island arc tholeiites" and abyssal tholeiites suggest that both have a common source in the upper mantle. In this model, similar to that proposed by Donnelly (1968), "island arc tholeiites" are partial melting products of the over-lying upper mantle beneath which "lithosphere" is thrust, whereas subsequent calcalkaline and shoshonite volcanism indicate partial melting of the down-going crust itself.

The former alternative is considered more likely for the following reasons:

1. Ni and Cr contents of Fiji tholeiites are 0—6 ppm and 0—15 ppm respectively, an order of magnitude less than abyssal examples (Engel *et al.*, 1965). This severely restricts the amount of possible olivine in their source areas.

2. The relative constancy of initial Sr 87/86 ratios in rocks of all periods suggests their corporate generation from source areas with comparable Rb—Sr histories. If the "island arc tholeiites" were derived from the upper mantle and subsequent volcanics from descending oceanic crust, a bi-model Sr isotope distribution would be expected. That Viti Levu samples have more radiogenic strontium than abyssal tholeiites is consistent with the former's two-stage derivation.

3. Although there are marked geochemical discontinuities between rocks of the first and second periods on Viti Levu, a broad continuum of iron enrichment and K_2O—SiO_2 relationships seems to exist when other areas are compared (Table 4 and Fig. 8). Although the similarity between "island arc tholeiites" of different regions is remarkable considering their corporate dissimilarity to typical calcalkaline suites, the overlap between the two is taken as evidence of a common source for both.

If Viti Levu is analogous with currently active island arcs, as this discussion has argued, then its southern coast was probably bordered by an active oceanic trench at least until the end of its second period of activity, or Tertiary *g* stage (5—10 m.y. ago; Page and McDougall, in prep.). If shoshonite volcanism is also genetically related to underthrust "lithosphere", this proximity to an active trench system persisted until 4—5 m.y. ago. For the southern coast of Viti Levu to have once faced an active direction of southwest Pacific sea floor spreading requires that it has since drifted or been rotated to its present position. Paleontological evidence also suggests that Viti Levu and some New Hebrides islands were once closer together (Coleman, 1969).

Thus, Viti Levu has at least a two-fold importance. Its igneous rocks offer a well-preserved and rather complete record of island arc evolution, and its apparent

relationships with an active boundary of sea floor spreading place one more constraint on geological reconstructions of the enigmatic southwest Pacific.

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Appendix 1. Trace Element Concentrations Used and Obtained for Interlaboratory Standards

Parenthese indicate values obtained when synthetic standards (**) or (G-1(*) were used for calibration. All concentrations are in ppm.

	G-1	W-1	AGV-1	GSP-2	G-2	BCR-1	S-1	PCC-1	DTS-1
Cs	1.5	0.9							
Rb	210	(22)*							
Ba	1,030	(192)*							
Sr	247	(184)*	(634)*						
Pb	49	(8.6)*							
Th	50	2.2	(6.2)**	(106)**	(25)**	(6.1)**			
U	3.2	0.54	(1.8)**	(2.4)**	(2.2)**	(1.6)**			
Zr	210	(91)*							
Hf	6.0	2.0							
Sn	3.0	2.5							
La	100	11.9							
Ce	170	20							
Pr	17	3.5							
Nd	55	12.5							
Sm	9.0	3.4							
Eu	1.4	1.1							
Gd	6.9	3.0							
Tb	0.70	0.60							
Dy	5.3	4.0							
Ho	0.5	0.8							
Er	1.4	2.08							
Tm	0.2	0.28							
Yb	1.0	2.1							
Y	13	(22)*							
Cu			61	33		23		10	7
Ni		76				11			
Co	2.3	47	16	7.0	4.6	38	18		145
V		245	116	60	40	360		32	
Sc	2.8	34	12	6.0	3.2	32	15	7.4	3.0
Cr	20	115	10	14	8.0	15	52		
Sb	0.30	0.96							

Appendix 2. Sample Locations and Descriptions

(Sample localities followed by T numbers refer to sites discussed by Tarling (1967) and apply to samples upon which he undertook paleomagnetic studies.)

68-48 Wainimala Group, Numbuonamboto Volcanic Conglomerate. E 177° 55′, S 18° 15′; Queen's Rd. (T 8). Flow. Phenocrysts: labradorite with strong zoning (30%), augite (<1%) replaced by chlorite + calcite. Matrix (70%): plagioclase, opaque, chlorite, calcite.

68-49 Wainimala Group, Numbuonamboto Volcanic Conglomerate. E 177° 43′, S 18° 12′; Queen's Rd. (T 17). Vesicular flow. Phenocrysts: turbid plagioclase with original zoning preserved (40%); fresh augite (10%). Matrix (50%): plagioclase, opaque, pyroxene, calcite.

68-50 Wainimala Group, Numbuonamboto Volcanic Conglomerate. E 177° 42′, S 18° 12′; Queen's Rd. (T 20). Flow. Phenocrysts: turbid plagioclase with original zoning preserved (20%), pyroxene replaced by chlorite + calcite + opaque (2%). Groundmass (78%): plagioclase, opaque, calcite, chlorite, relict pyroxene.

68-51 Wainimala Group, Numbuonamboto Volcanic Conglomerate. Same site as 68-50. Dike or recrystallized flow; fine-grained, holocrystalline texture. Plagioclase, chlorite after hornblende, quartz, minor calcite.

68-52 Wainimala Group, Numbuonamboto Volcanic Conglomerate E 177° 35′, S 18° 11′; Queen's Rd. (T 22). Aphanitic flow, trachytic texture. Plagioclase laths with interstitial chlorite, opaque, and calcite.

68-54 Wainimala Group, undifferentiated. E 177° 33′, S 18° 03′; Singatoka Valley Rd. (T 32). Flow. Phenocrysts: turbid plagioclase with original oscillatory zoning preserved (10%); hornblende (<1%). Matrix (90%): plagioclase, opaque, calcite, chlorite.

68-55 Wainimala Group, undifferentiated. E 177° 23′, S 18° 05′; Queen's Rd. (T 36). Flow. Phenocrysts: altered plagioclase (3%); pyroxene altered to chlorite + epidote + calcite (2%). Matrix (95%): plagioclase, opaque, chlorite, calcite.

68-56 Tholo Plutonics. E 177° 32′, S 18° 03′; Nakalavo village. Dike. Subhedral, complexly zoned labradorite (70%); augite (13%); opaque (2%); chlorite, epidote, muscovite. calcite (15%).

68-57 Tholo Plutonics. Same site as 68-56. Dike. Subhedral, zoned labradorite (65%); zoned augite (15%); chlorite, epidote, muscovite, calcite (20%).

68-58 Tholo Plutonics. E 177° 32′, S 17° 50′. Yavuna village (T 41). Dike. Zoned labradorite often with feldspar-quartz intergrowths and chloritic alterations (40%); augite, some altered to chlorite + calcite (20%); quartz (15%); opaque (2%); chlorite (15%); epidote, sphene, calcite (8%).

68-59 Tholo Plutonics. Same site as 68-56. Tonalite. Zoned, altered plagioclase, with ubiquitous calcite and epidote (50%); pyroxene altered to chlorite + opaque (2%); quartz (40%); epidote, chlorite, calcite, opaque (8%).

68-60 Savura Volcanic Group, Nasinu Basalt. E 178° 29′, S 18° 04′; Prince's Rd. Quarry (T 3). Aphanitic flow. Plagioclase, devitrified glass, pyroxene (usually altered), and opaque.

68-61 Savura Volcanic Group, Nasinu Basalt. E 178° 30′, S 18° 05′; P.W.D. Nasinu Quarry (T 2). Vesicular flow. Trachytic textures. Two generations of plagioclase (85%) and pyroxene (10%). Pyroxene is often oxidized with clusters of opaques.

68-62 Savura Volcanic Group, Nasinu Basalt. Same site as 68-61 and with similar mineralogy.

68-63 Koroimavua Andesitic Group. E 177° 27′, S 17° 48′; P.W.D. Lomolomo Quarry. Clast from volcanic conglomerate. Phenocrysts: zoned labradorite (30%); pale-green augite (15%); biotite (5%). Matrix (50%): plagioclase, pyroxene, opaque. Accessory apatite.

68-64 Mba Basaltic Group, undifferentiated. E 177° 26′, S 17° 40′; Varangge Crk. (T 45). Flow. Phenocrysts: labradorite (25%); olivine (2%). Matrix (73%): plagioclase microlites, pyroxene, opaque.

68-65 Mba Basaltic Group, undifferentiated. Same site as 68-64. Flow. Phenocrysts: labradorite (35%); olivine (5%); opaque (2%). Matrix (48%): plagioclase microlites, pyroxene, opaque.

68-66 Andesite plug. E 178° 14′, S 18° 11′; Mau Quarry. Phenocrysts: zoned labradorite (40%); hornblende (20%). Matrix (40%): plagioclase, pyroxene, opaque.

68-71 Koroimavua Andesitic Group. E 177° 34′, S 17° 49′; abandoned quarry, Nausori Rd. Clast from volcanic conglomerate. Phenocrysts: zoned labradorite (20%); zoned pale-green augite (30%), olivine pseudomorphs (2%). Matrix (48%): plagioclase, pyroxene, opaque.

68-72 Mba Basaltic Group, undifferentiated. E 177° 38′, S 17° 34′; Nandarivatu Rd. (T 50). Flow. Phenocrysts: pale green zoned augite (25%); olivine usually altered (8%). Matrix (67%): plagioclase microlites, pyroxene, opaque.

68-73 Mba Basaltic Group, undifferentiated. E 177° 37′, S 17° 33′; Nandarivatu Rd. (T 54). Flow. Phenocrysts: labradorite (10%); pale-green augite (5%), olivine pseudomorphs (1%). Matrix (84%): Plagioclase microlite, pyroxene, opaque.

68-74 Mba Basaltic Group, Nakorotumbu Basalt. E 178° 15′, S 17° 22′; King's Rd. (T 58). Flow. Phenocrysts: zoned augite (40%); olivine, partially altered (20%). Matrix (40%): plagioclase, pyroxene, resorbed olivine.

68-75 Mba Basaltic Group, undifferentiated. E 177° 31′, S 17° 33′. King's Rd. (T 46). Flow. Phenocrysts: plagioclase (35%), augite (25%), olivine pseudomorphs (5%), opaques (2%). Matrix (33%): plagioclase, pyroxene, opaque.

68-76 Mba Basaltic Group, undifferentiated. E 177° 46′, S 17° 27′; King's Rd. (T 49). Flow. Phenocrysts: zoned labradorite (50%); augite (3%); olivine pseudomorphs (1%). Matrix: plagioclase, pyroxene, opaque, sanidine (?).

References

Ahrens, L. H., Taylor, S. R.: Spectrochemical Analysis, 2nd ed. Reading, Mass: Addison-Wesley 1961.

Arriens, P. A., Compston, W.: A method for isotopic ratio measurement by voltage peak switching, and its application with digital output. J. Mass Spectrometry and Ion Physics **1**, 471—481 (1968).

Baker, P. E.: Petrology of Mt. Misery volcano, St. Kitts, West Indies. Lithos **1**, 124—150 (1968a).

— Comparative volcanology and petrology of the Atlantic island-arcs. Bull. Volcanol. **32**, 189—206 (1968b).

Band, R. B.: The geology of Southern Viti Levu and Mbengga. Bull. Geol. Surv. Fiji **15** (1968).

Banghar, A. R., Sykes, L. R.: Focal mechanisms of earthquakes in the Indian Ocean and adjacent regions. J. Geophys. Res. **74**, 632—650 (1969).

Bartholomew, R. W.: Geology of the Lautoka Area, Northwest Viti Levu. Bull. Geol. Surv. Fiji **2** (1959).

— Geology of the Nandi area, Western Viti Levu. Bull. Geol. Surv. Fiji **7** (1960).

Blake, D. H.: Post Miocene volcanoes on Bougainville Island, Territory of Papua and New Guinea. Bull. Volcanol. **32**, 121—138 (1968).

Byers, F. M., Jr.: Petrology of three volcanic suites, Umnak and Bogoslof Islands, Aleutian Islands, Alaska. Bull. Geol. Soc. Am. **72**, 93—128 (1961).

Cann, J. R.: Spilites from the Carlsberg Ridge, Indian Ocean. J. Petrology **10**, 1—19 (1969).

Coats, R. R.: Magmatic differentiation in Tertiary and Quaternary volcanic rocks from Adak and Kanaga Islands, Aleutian Islands, Alaska. Bull. Geol. Soc. Am. **63**, 485—514 (1952).

— Basaltic andesites. In: Basalts (ed. H. H. Hess and A. Poldevaart), vol. **2**, p. 689—736. New York: Interscience Publishers 1968.

Cole, W. S.: Upper Eocene and Oligocene larger foraminifera from Viti Levu, Fiji. U.S. Geol. Surv. Profess. Papers **374-A** (1960).

Coleman, P. J.: Derived Eocene larger foraminifera on Maewo, Eastern New Hebrides and their south-west Pacific implications. 1967 Ann. Report New Hebrides Geol. Surv., p. 36—37 (1969).

Compston, W., Lovering, J. F., Vernon, M. J.: Rubidium-strontium age of the Bishopville aubrite and its component enstatite and feldspar. Geochim. Cosmochim. Acta **29**, 1085—1099 (1965).

Cooper, J. A.: The flame photometric determination of potassium in geological materials used for potassium-argon dating. Geochim. Cosmochim. Acta **27**, 525—546 (1963).

Crook, K. A.: Burial metamorphic rocks from Fiji. New Zealand J. Geol. Geophys. **6**, 681—704 (1963).

Cullen, D. J.: Island arc development in the southwest Pacific. Tectonophysics **4**, 163—172 (1967).

Dickinson, W. R.: Tectonic Development of Fiji. Tectonophysics **4**, 543—553 (1967).

— Sedimentation of volcaniclastic strata of the Pliocene Koroimavua group in northwest Viti Levu, Fiji. Am. J. Sci. **266**, 440—453 (1968a).

— Circum-Pacific andesite types. J. Geophys. Res. **73**, 2261—2269 (1968b).

— Rickard, M. J., Coulson, F. I., Smith, J. G., Lawrence, R. L.: Late Cenozoic shoshonitic lavas in northwestern Viti Levu, Fiji. Nature **219**, 148 (1968).

Donnelly, T. W.: Caribbean island-arcs in light of the sea-floor spreading hypothesis. Trans. N.Y. Acad. Sci. **30**, 745—750 (1968).

Rogers, J. J. W., Pushkar, P., Armstrong, R. L.: Chemical evolution of the igneous rocks of the Eastern West Indies: an investigation of Thorium, Uranium, and Potassium distributions, and lead and strontium isotopic ratios (in prep.).

Engel, A. E. J., Engel, C. G., Havens, R. G.: Chemical characteristics of oceanic basalts and the upper mantle. Bull. Geol. Soc. Am. **76**, 719—734 (1965).

Frey, F. A., Haskin, M. A., Pbetz, J. A., Haskin, L. A.: Rare earth abundances in some basic rocks. J. Geophys. Res. **73**, 6085—6098 (1968).

Gast, P. W.: Trace element fractionation and the origin of tholeiitic and alkaline magma types. Geochim. Cosmochim. Acta **32**, 1057—1086 (1968).

Gorshkov, G. S.: Catalogue of the active volcanoes of the world. Part 7, Kurile Islands. Int. Assoc. Volcanology (1958).

— Petrochemistry of Kurile-Kamchatka Cenozoic volcanic province [in Russian], Moscow, 277 pp. (1966).

Graham, A. L., Nicholls, G. D.: Mass spectrographic determinations of lanthanide element contents in basalts. Geochim. Cosmochim. Acta **33**, 555—568 (1969).

Green, T. H., Ringwood, A. E.: Genesis of the calc-alkaline igneous rock suite. Contr. Mineral. and Petrol. **10**, 105—162 (1968).

Hart, S. R.: K, Rb, Cs contents and K/Rb, K/Cs ratios of fresh and altered submarine basalts. Earth Planet. Sci. Letters **6**, 295—303 (1969).

Hatherton, T., Dickinson, W. R.: The relationship between andesitic volcanism and seismicity in Indonesia, the Lesser Antilles, and other island arcs. J. Geophys. Res. **74**, 5301—5310 (1969).

Heier, K. S., Rogers, J. J. W.: Radiometric determination of thorium, uranium, and potassium in basalts and in two magmatic differentiation series. Geochim. Cosmochim. Acta **27**, 137—154 (1963).

Hirst, J. A.: Geology of East and Northeast Viti Levu. Bull. Geol. Surv. Fiji **12** (1965).

Houtz, R. E.: Geology of Singatoka area, Viti Levu. Bull. Geol. Surv. Fiji **6** (1960).

— Regional geology — Keiyasi area. Bull. Geol. Surv. Fiji **10** (1963).

Hubbard, N. J.: A chemical comparison of oceanic ridge, Hawaiian tholeiitic, and Hawaiian alkalic basalts. Earth Planet. Sci. Letters **5**, 346—352 (1969).

Ibbotson, P.: Geology of the Suva area, Viti Levu. Bull. Geol. Surv. Fiji **4** (1960).

— Petrological aspects of some volcanic rocks on Fiji. Unpub. Ph. D. thesis, Univ. London (1963).

— Petrology of the Tertiary Caldera, Tavua goldfield. Mem. Geol. Surv. Fiji **3** (1967).

Isacks, B., Oliver, J., Sykes, L. R.: Seismology and the New Global Tectonics. J. Geophys. Res. **73**, 5855—5899 (1968).

— — — Focal mechanisms of deep and shallow earth-quakes in the Tonga-Kermadec region and the tectonics of island arcs. Bull. Geol. Soc. Am. **80**, 1443—1470 (1969).

Isshiki, N.: Petrology of Hachijo-Jima volcanic group, seven Izu Islands, Japan. J. Fac. Sci. Univ. Tokyo **15**, 91—134 (1963).

Jakes, P., Gill, J. B.: Rare earth elements and island arc tholeiites (in prep.).
— White, A. J. R.: Structure of the Melanesian Arcs and correlation with distribution of magma types. Tectonophysics 8, 223—236 (1970).
— — K/Rb ratios of rocks from island arcs (in prep.).
Joplin, G. A.: The shoshonite association: a review. J. Geol. Soc. Australia **15**, 275—294 (1968).
Karig, D. E.: Ridges and basins of the Tonga-Kermackc Island Arc system. J. Geophys. Res. **75**, 239—254 (1970).
— Kermadec arc — New Zealand tectonic confluence (in prep.).
Katsui, Y.: Andesites from the Andes and Antarctica. Proc. Symposium on Andesites, Univ. Oregon (in press) (1969).
Kuno, H.: Petrology of Hakone volcano and the adjacent areas, Japan. Bull. Geol. Soc. Am. **61**, 957—1020 (1950).
— Catalogue of the active volcanoes of the world. Part 11, Japan, Taiwan and Marianas. Int. Assoc. Volcanology (1962).
— Lateral variation of basalt magma across continental margins and island arcs. Bull. Volcanol. **29**, 195—222 (1966).
Ladd, H. S.: Geology of Viti Levu, Fiji. Bernice P. Bishop Mus. Bull. **119** (1934).
Lowder, G. G., Carmichael, I. S. E.: The volcanoes and caldera of Talasea, New Britain. Bull. Geol. Soc. Am. **81**, 17—38 (1970).
MacDonald, G. A., Katsura, T.: Chemical composition of Hawaiian lavas. J. Petrology **5**, 82—133 (1964).
McBirney, A. R., Aoki, K.: Petrology of the island of Tahiti. Geol. Soc. Am. Mem. **116**, 523—556 (1968).
McDougall, I.: K-Ar ages of some rocks from Viti Levu, Fiji. Nature **198**, 677 (1963).
Masuda, A.: Lanthanides in basalts of Japan with three distinct types. Geochem. J. **1**, 11—26 (1966).
— Geochemistry of lanthanides in basalts of Central Japan. Earth Planet. Sci. Letters **4**, 284—292 (1968).
Melson, W. G., Andel, T. van: Metamorphism in the Mid-Atlantic Ridge, 22° N Latitude. Marine Geol. **4**, 165—186 (1966).
— — Thompson, G.: Volcanism and metamorphism in the mid-Atlantic Ridge, 22° N Latitude. J. Geophys. Res. **73**, 5925—5943 (1968).
Mitchell, A. H. G.: Geology of South Malekula. New Hebrides Geol. Surv. Rep. **3** (1966).
Morgan, W. R.: A note on the petrology of some lava types from East New Guinea. J. Geol. Soc. Australia **13**, 583—591 (1966).
Nockolds, S. R.: Average chemical compositions of some igneous rocks. Bull. Geol. Soc. Am. **65**, 1007—1032 (1954).
— Allen, R.: The geochemistry of some igneous rock series. Geochim. Cosmochim. Acta **4**, 105—142 (1953).
Norrish, K., Chappell, B. W.: X-ray fluorescence spectrography. In: Physical methods in determinative mineralogy (ed. J. Zussman), p. 161—214. London: Academic Press 1967.
Page, R., McDougall, I.: Potassium-Argon dating of the Tertiary f_{1-2} Stage in New Guinea and its bearing on the geological time-scale (in prep.).
Richard, J. J.: Catalogue of the active volcanoes of the world, Part 13, Kermadec, Tonga and Samoa. Int. Assoc. Volcanology (1962).
Rickard, M. J.: The geology of Mbalevuto area. Bull. Geol. Surv. Fiji **11** (1963).
— Reconnaissance geology of Vanua Levu. Fiji Geol. Surv. Mem. **2** (1966).
Robertson, E. I.: Bouguer anomaly map of Viti Levu, Fiji. New Zealand J. Geol. Geophys. **10**, 1309—1312 (1967).
Rodda, P.: Outline of the geology of Viti Levu. New Zealand J. Geol. Geophys. **10**, 1260—1273 (1967).
— Analyses of rocks from Fiji. Suva: Oceania Printery. 45 pp. (1969).
— Band, R. B.: Geology of Viti Levu (map). Hobart: Mercury Press (1966).
— Snelling, N. J., Rex, D. C.: Radiometric age data on rocks from Viti Levu, Fiji. New Zealand J. Geol. Geophys. **10**, 1248—1259 (1967).
Schmidt, R. G.: Petrology of the volcanic rocks, Saipan, Mariana Islands. U.S. Geol. Surv. Profess. Papers **280**, 127—176 (1957).

Sclater, J. G., Menard, H. W.: Topography and heat flow of the Fijian Plateau. Nature **216**, 991—993 (1967).
Scripps Inst. of Oceanography: Bathymetry of the Southwest Pacific (chart) (1968).
Seeley, J. B.: Geology of Raki Raki area, Viti Levu, Fiji. Unpub. M. Sc. Thesis, Univ. Auckland (1967).
Shor, G. G., Kirk, H. K., Menard, H. W.: Crustal structure of the Melanesian area (in prep.).
Skiba, W. J.: Geological studies in Southwest Viti Levu. Mem. Geol. Surv. Fiji **1** (1964).
Smith, R. E.: Redistribution of major elements in the alteration of some basic lavas during burial metamorphism. J. Petrology **9**, 191—219 (1968).
Solomon, S., Biehler, S.: Crustal structure from gravity anomalies in the southwest Pacific. J. Geophys. Res. **74**, 6696—6701 (1969).
Stark, J. T.: Petrology of the volcanic rocks of Guam. U.S. Geol. Surv. Profess. Papers **403-C** (1963).
Summerhayes, C. P.: Note on Macquarie ridge and the Tonga-Kermadec complex. New Zealand J. Sci. **10**, 808—812 (1967).
Sykes, L. R., Isacks, B. L., Oliver, J.: Spatial distributiin of deep and shallow earthquakes of small magnitude in the Fiji-Tonga region. Bull. Seism. Soc. Am. **59**, 1093—1114 (1968).
Tarling, D. H.: The paleomagnetism of some rock samples from Viti Levu, Fiji. New Zealand J. Geol. Geophys. **10**, 1235—1247 (1967).
Tatsumoto, M.: Isotopic composition of lead in volcanic rocks from Hawaii, Iwo Jima, and Japan. J. Geophys. Res. **71**, 1721—1733 (1966).
— Hedge, C. E., Engel, A. E. J.: Potassium, rubidium, strontium, thorium, uranium, and the ratio of Sr-87 to Sr-86 in oceanic tholeiitic basalt. Science **150**, 886—888 (1965).
Taylor, S. R.: Geochemical analysis by spark source mass spectrography. Geochim. Cosmochim. Acta **29**, 1243—1261 (1965).
— Geochemistry of andesites. In: Origin and distribution of the elements (ed. L. H. Ahrens), p. 559—583. Oxford: Pergamon Press 1968.
— Capp, A. C., Graham, A. L., Blake, D. H.: Trace element abundances in andesites. II Saipan, Bougainville and Fiji. Contr. Mineral. and Petrol. **23**, 1—26 (1969).
— White, A. J. R.: Trace element abundances in andesites. Bull. Volcanol. **29**, 177—194 (1966).
Vallance, T. G.: Concerning spilites. Proc. Linnean Soc. N.S. Wales **85**, 8—52 (1960).
— On the chemistry of pillow lavas and the origin of spilites. Mineral. Mag. **34**, 471—481 (1965).
Warden, A. J.: The geology of the Central Islands. New Hebrides Geol. Surv. Rep. **5** (1967).
Wilcox, R. E.: Igneous rocks of the Near Islands, Aleutian Islands, Alaska. Int. Geol. Congr. XX Sess., Mexico **11-A**, 365—378 (1959).
Wilson, J. T.: A new class of faults and their bearing on continental drift. Nature **207**, 343—347 (1965).
Woolnough, W. G.: The continental origin of Fiji. Proc. Linnean Soc. N.S. Wales **28**, 457—540 (1903).

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RARE EARTH ELEMENTS AND THE ISLAND ARC THOLEIITIC SERIES

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pp. 17 – 28

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The authors wish to make the following corrections :

p. 17 Abstract: line 4. "... rocks with these characteristics occur above shallower earthquake foci ...".

p. 17 Add to bottom of column two "and alkali basalts or shoshonites afterwards".

p. 20 Column two, 1st line of 3rd paragraph under Results. "Fig. 6 summarizes published island arc REE data ...".

p. 24 Column two, line 4. "... it may yet prove correct, it is at least unnecessary to ..."

p. 25 Column one, 7th line of 2nd paragraph under Genesis. "... calc-alkaline series because ..."

p. 25 Lines 16-17 of above paragraph.
"Calc-alkaline magmatism (s.s.) in oceanic island arcs may alternatively result from partial melting of ..."

Although this occurrence of tholeiitic rocks in island arcs was long emphasized by Kuno [11] and recognized by Tilley [12], it seems to have been largely ignored or thought volumetrically insignificant – at most, one parent from which a calc-alkaline series might have differentiated. We contend, however, not only that there is a complete and separate "island arc tholeiitic magma series" which can be distinguished chemically, spatially, and temporally from the "calc-alkaline magma series", but also that it is at least as if not more common than the calc-alkaline series in Tertiary circum-Pacific island arcs. In this paper we summarize evidence for this contention and present rare element data for some Recent and older island arc tholeiites. Using these data we emphasize the differences between rocks of the island arc tholeiitic and calc-alkaline series, re-emphasize that abyssal tholeiites are not the only terrestrial rocks with condritic rare earth abundance patterns, and argue that the ophiolite complexes of island arcs and orogenic areas need not be considered fossil remnants of ocean crust.

2. Comparison between the calc-alkaline and island arc tholeiitic series

The distinctive rocks of island arcs and orogenic zones are an association which on petrographic criteria would be called basalt, andesite, dacite, and rhyolite, in which intermediate members are more common than basic. Such an association is usually called the "calc-alkaline series". It is both possible and useful, however, to distinguish between two different rock series in many island arcs both having these general petrographic characteristics. (A third and alkaline series also occurs but will not be considered here.) Failure to make this distinction has led to confusion about the nature of island arc magmatism and added unnecessary complexity to the "andesite problem".

Petrographically the "typical" calc-alkaline series is principally characterized by large plagioclase phenocrysts with complex oscillatory zoning, moth-eaten appearance, and calcic cores, and by groundmass pyroxene which is more likely to be hypersthene than pigeonite. Chemically its most distinguishing features are a silica mode of 58-59%, lack of iron enrichment (fig. 1), and moderate concentrations and enrichment of K_2O (fig. 2); other useful chemical features are summarized in table 1.

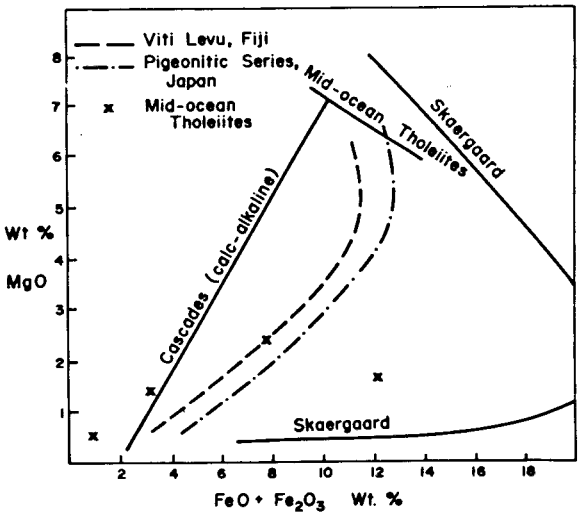

Fig. 1. Generalized MgO-FeO+Fe_2O_3 relationships in selected igneous provinces. Data for the Cascades are from [54], Fiji [8], pigeonitic series of Japan [11], mid-ocean tholeiites [23], and the Skaergaard [55].

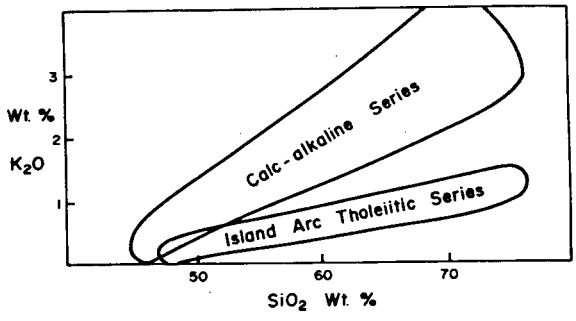

Fig. 2. Generalized K_2O-SiO_2 relationships in selected island arcs. Data from Gill [8], (fig. 8). Analyses from Tonga, Fiji, S. Kurile and S. Sandwich islands, Saipan-Guam, and the pigeonitic series of Japan and the Izu-N. Mariana islands all fall within the island arc tholeiitic series field. Analyses from the Aleutian and N. Kurile islands, central New Guinea, Bougainville, the New Hebrides, and central Viti Levu, Fiji fall within the calc-alkaline series field. Note that analyses from St. Kitts, Lesser Antilles also plot within the island arc tholeiitic field but have no iron-enrichment. Similarly, those from northern New Guinea and New Britain have characteristic iron enrichment but nevertheless plot within the calc-alkaline field.

Table 1
Comparison of means and ranges for selected chemical features of three rock series.

	Calc-Alkaline Series	Island Arc Tholeiitic Series	Abyssal Tholeiitic Series
SiO_2 Range	53-70%	45-70%	47-62%
mode	59%	53%	49%
TiO_2	0.5-1.2%	0.5-1.5%	1.0-2.5%
Al_2O_3	16-19%	14-19%	14-19%
Na_2O/K_2O	2-3	4-6	10-15
Rb	30 ppm	3-10 ppm	0.2-5.0 ppm
Sr	380 "	100-200 "	70-150 "
Ba	270 "	50-150 "	6-30 "
Pb	3-7 "	2-4 "	0.5 "
Cs	0.5-1.0 ppm	0.1 "	0.05 "
K/Rb	400-500	1000	1000
Rb/Sr	0.05-0.10	0.01-0.05	0.02
Th	2 ppm	0.5 ppm	0.15 ppm
U	0.7 "	0.3 "	0.10 "
Th/U	3-4	1-2	1-2
Ni	18 "	0-30 "	30-200 ppm
Cr	56 "	0-50 "	200-400 "
La	12 "	1-6 "	2-8 "
La/Yb	6-8	1-2	1-2

Data for the calc-alkaline series are compiled principally from this paper, Nockolds and Allen [3], Taylor [4], Chayes [48] and Taylor et al. [20]; data for the island arc tholeiitic series are from this paper, Nockolds and Allen [8], Heier and Rogers [49], Masuda [18], Tatsumoto [37], Gill [8], Donnelly et al. [9], Ewart (unpublished) and Gill (unpublished); data for the abyssal tholeiitic series are from Engel et al. [50], Frey et al. [19], Tatsumoto [37], Philpotts et al. [51], Hart [52], Aumento [53] and Kay et al. [23].

Also in table 1 are data for many rocks occurring in island arcs which it is chemically inappropriate to call calc-alkaline. Although they have a similar silica range, the mode is 53-54%. They have varying degrees of iron enrichment (fig. 1), less K_2O (fig. 2) and Rb, Ba, Cs, Pb, and Sr, higher Na_2O/K_2O and K/Rb ratios, lower Th and U contents and Th/U ratios, low Ni (0-30 ppm) and Cr (0-50 ppm), and REE concentrations as discussed later. Analyses of samples with intermediate silica content from each series are contrasted in table 2. Spatially these rocks occur closer to the trench (i.e., above shallower earthquake foci) than do the calc-alkaline. Temporally they represent the earliest volcanism in oceanic island arc history and often form an extensive basement upon which subsequent calc-alkaline volcanism develops. Their phenocryst mineralogy is similar to that of calc-alkaline rocks but phenocrysts are less abundant, there are more opaques, and the groundmass pyroxene is more likely to be pigeonite than hypersthene. These can be distinguished from calc-alkaline rocks in many Tertiary [7,8], Paleozoic [14] and even Archean orogenic zones [15].

Because these rocks are not calc-alkaline as traditionally understood, and because they persistently differ in chemistry, space, and time from those which are calc-alkaline, we propose to emphasize these differences by using a name other than calc-alkaline to describe them. Because they are less potassic, more iron-enriched, more likely to have groundmass pigeonite in their basic members, and were considered tholeiitic by Tilley [12], we suggest that they be called the "island arc tholeiitic series". "Tholeiitic" is thus used to describe rocks with silica contents ranging between 45-75%. As in any empirical classification the criteria employed are relative and the classes are end members of a continuum between which there is considerable overlap. Although in any one area (i.e., Fiji,

Table 2
Comparison of intermediate rocks from the calc-alkaline and island arc tholeiitic series of Viti Levu, Fiji.

	Calc-alkaline series*	Island arc tholeiitic series**
SiO_2	57.4	57.4
TiO_2	0.65	1.25
Al_2O_3	17.7	15.6
Fe_2O_3	3.20	3.48
FeO	3.30	5.01
MgO	3.45	3.38
CaO	7.40	6.14
Na_2O	3.40	4.20
K_2O	1.25	0.43
P_2O_5	-	0.44
Rb	22	4.0
Ba	270	50
Cs	0.47	<0.1
Pb	2.2	1.2
Sr	540	219
K/Rb	472	890
Rb/Sr	0.04	0.01
Th	1.4	0.31
U	0.29	0.34
Th/U	4.8	0.9
La	9.1	2.4
La/Yb	10.3	1.0
Ni	8	nd
Cr	36	2
Sr 87/86	0.7040	0.7037

nd = not detectable; - = not measured
* Data from Taylor et al. [20]; Table 2, sample X-96 except Ni and Cr data from sample X-88 and Sr 87/86 for X-88 (Gill, unpublished).
** Data from Gill [8]; Table 1, sample 68-55 except for Cs, La and Yb data from sample 68-60.

the Lesser Antilles) there be no overlap, some of these criteria and particularly the amount of iron enrichment may vary independently thereby blurring the distinction.

3. Analytical methods

REE data were obtained using spark-source mass spectrographic methods similar to those of Taylor [16]. Lu 175 was used as an internal standard and the W-1 and G-1 concentrations adopted by Gill [8] were used for calibration except for Gd where his values were doubled. Samples were analyzed in duplicate or triplicate. Precision, expressed as relative deviation (1σ) and measured by replicate analyses, is ± 10% for all elements except Gd which is ± 15%.

4. Results

Flat rare earth abundance patterns parallel but enriched relative to those of chondrites are thought by some [17] to unambiguously characterise abyssal basalts of mid-ocean ridges. Such patterns have been known in rocks of islands arcs for some time [18,19], but, as with the tholeiitic volcanics in which they occur, they have been largely ignored.

Rare earth data for nineteen samples are listed in table 3 and presented relative to the average of nine chondrites [19] in figs. 3-5. These samples were chosen to represent the variety of rocks occurring in several southwest Pacific island arcs from which we had specimens. "Chondritic" patterns are shown in each figure characterizing samples from Recent volcanism in New Britain, basement complexes of Eastern Papua, New Guinea, the Solomon islands, and Macquarie Ridge, two Archean and one Proterozoic site. These flat rare earth patterns are characteristic of rocks of the islands arc tholeiitic series. They have been described in rocks of this series from Japan [18], Saipan [20], Fiji [8], and Tonga [21].

Fig. 6 summarizes published islands arc REE data (54 analyses from 9 western Pacific arcs) and compares them with data for mid-ocean tholeiites [22]. Rare earth abundance patterns in abyssal and arc tholeiitic series are almost identical and no more characteristic of one than the other. Whatever genetic constraints are implied by such patterns, they apply equally to both. No significant positive or negative Eu anomalies occur in our tholeiitic samples (although Gd precision is poor) which suggests that their Al_2O_3 contents are primary and not the accidental result of plagioclase accumulation, and that plagioclase crystallization has not been sufficient to affect the REE distribution as is argued for abyssal tholeiites [23].

From figs. 3-6 it is clear that rocks of the calc-alkaline and island arc tholeiitic series have different rare earth abundance patterns, a difference that is

Table 3
New data on rare earth elements in island arc rocks and older orogenic areas.

	Solomon Islands			Eastern Papua				Macquarie Island		Norseman, West Australia	
	tholeiites			tholeiites						Greenstones	
	7913	6640	6719	6785	2041	2207	3091	26	48	23759	23745
La	3.3	5.2	10.8	2.3	3.50	4.0	6.6	4.2	5.3	2.9	4.8
Ce	9.8	9.9	19.2	4.0	8.4	10.0	10.5	9.1	6.7	6.5	8.8
Pr	1.4	1.9	3.5	0.92	1.4	1.4	1.9	1.6	1.9	1.2	1.5
Nd	6.6	9.6	14.9	4.2	6.2	6.4	8.1	7.3	9.1	5.6	7.5
Sm	3.1	3.2	nd	1.2	2.4	2.4	3.1	2.7	2.8	nd	nd
Eu	0.91	1.1	1.4	0.53	0.84	0.76	0.99	1.0	0.95	0.73	0.95
Gd	4.1	4.2	4.0	1.6	3.4	3.2	4.6	4.8	3.8	1.5	2.2
Tb	0.5	0.58	0.75	0.26	0.42	0.46	0.51	0.48	0.45	0.41	0.34
Dy	4.5	4.3	6.7	2.7	3.8	3.8	5.3	4.5	4.7	3.8	3.0
Ho	1.1	0.91	1.3	0.55	0.80	0.87	1.2	1.0	0.98	0.80	0.70
Er	2.4	2.0	3.2	1.25	2.1	2.0	2.5	2.5	2.3	2.1	1.5
Tm	0.29	0.54	0.70	0.24	0.27	0.25	0.33	0.40	0.26	0.26	0.30
Yb	2.5	2.2	3.6	1.5	2.2	2.1	2.6	2.7	2.3	2.1	1.7
Ba	6.3	31.6	360.0	48.4	16.4	23.1	24.5	38.2	25.3	85.0	92.0

	Upper Proterozoic of Czech Massive spilite	New Britain		Guadalcanal		Eastern Papua	New Guinea Highlands	
		tholeiites		andesites			shoshonites	
	ZBECNO	5ING0089	5ING0098	144	160	3514B	105	101
La	2.8	1.1	2.5	8.8	10.6	36.1	13.8	18.4
Ce	6.9	2.6	5.8	15.9	15.8	61.5	28.6	34.8
Pr	0.90	0.44	0.81	nd	2.8	10.3	3.40	3.3
Nd	8.30	2.3	4.2	8.0	10.1	31.1	13.8	12.6
Sm	nd	1.0	1.6	1.4	nd	6.1	3.9	2.6
Eu	0.78	0.43	0.59	3.4	0.66	2.3	1.0	0.85
Gd	3.4	1.9	1.6	4.1	1.7	6.3	6.0	2.2
Tb	0.45	0.22	0.34	0.80	0.30	1.0	0.56	0.30
Dy	4.4	1.9	2.5	4.3	2.2	3.6	4.3	4.1
Ho	0.95	0.40	0.55	1.58	0.36	1.1	0.89	0.50
Er	2.1	1.0	1.5	2.6	0.98	2.4	2.0	1.2
Tm	0.56	0.20	0.31	0.30	0.15	0.5	0.24	0.2
Yb	2.1	1.4	1.3	3.1	1.0	3.0	2.1	1.2
Ba	44.5	62.1	96.0	300.0	1330.0	630.0	482.0	543.0

maintained even at SiO_2 = 70-75% [8,20]. In contrast to the chondritic patterns of island arc tholeiitic rocks, calc-alkaline samples have patterns enriched in the lighter elements (La-Sm) with concentrations similar to those of Hawaiian tholeiites [24]. They have La/Yb ratios of 5-20, clustering around 6-8, whereas those of the island arc tholeiitic series are less than 3 and usually 1-2. Calc-alkaline and more K_2O-rich rocks having these inflected REE patterns and higher La/Yb ratios have also been reported from Japan [18,25], Bougainville [20], Fiji [8,20], and New Zealand [25].

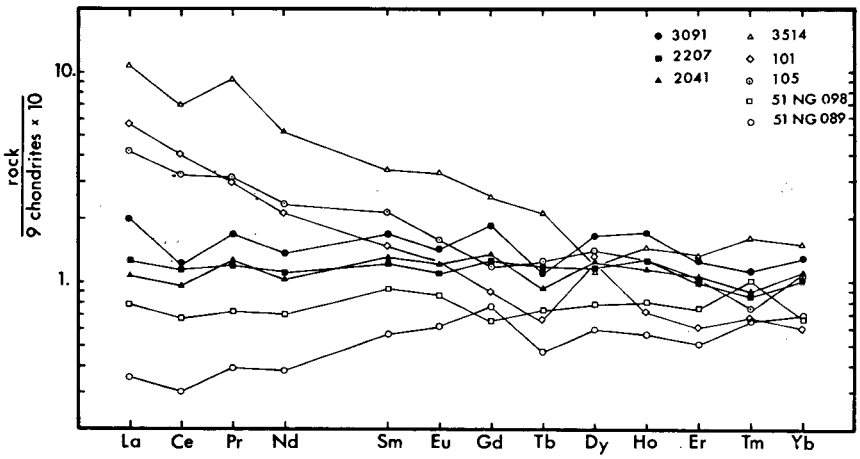

Fig. 3. Relative rare earth element patterns for rocks from Papua-New Guinea. Empty symbols Recent lavas: 3514 - Andesite from Mt. Victory, Cape Nelson-East Papua, 101 and 105 latite and shoshonite from New Guinea Highlands, Mt. Giluwe and Mt. Hagen respectively, 51 NG 098 and 51 NG 089 island are tholeiites from Ulawan volcano New Britain. Full symbols basement tholeiitic lavas from Eastern Papua (samples by courtesy of I. Smith).

There is a weak correlation between K_2O and La (fig. 7 and [20]) and between K_2O and La/Yb ratios (fig. 8) in island arcs. This may be significant in light of the apparent depth dependence of K_2O in such samples [26]. However, in contrast to Japanese alkali basalts, most New Guinea and Fiji shoshonites have neither the high La nor correspondingly high La/Yb ratios expected in high-K rocks.

It is of fundamental importance to realize that volcanic rocks with chondritic REE patterns are *characteristic* of the early tholeiitic stages in island arc evolution. The recent eruption of such rocks in Japan, New Britain, and Tonga and their occurrence in the mid-Tertiary basement complexes of Saipan and several Melanesian arcs make it unnecessary to assume that such similar rocks as the Mesozoic ophiolites of

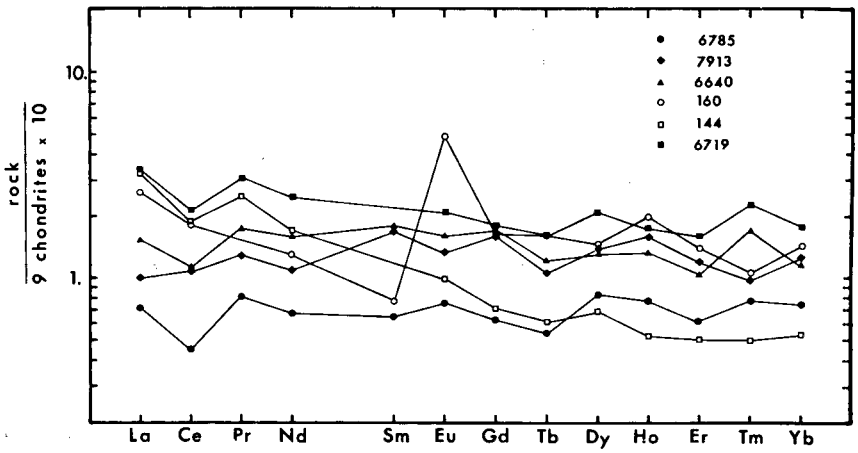

Fig. 4. Relative rare earth element patterns for rocks from Solomon Islands. Empty symbols Recent lavas from Savo (160) and Guadalcanal (144). Full symbols Tertiary lavas of Guadalcanal basement (samples by courtesy of Dr. B. Hackman).

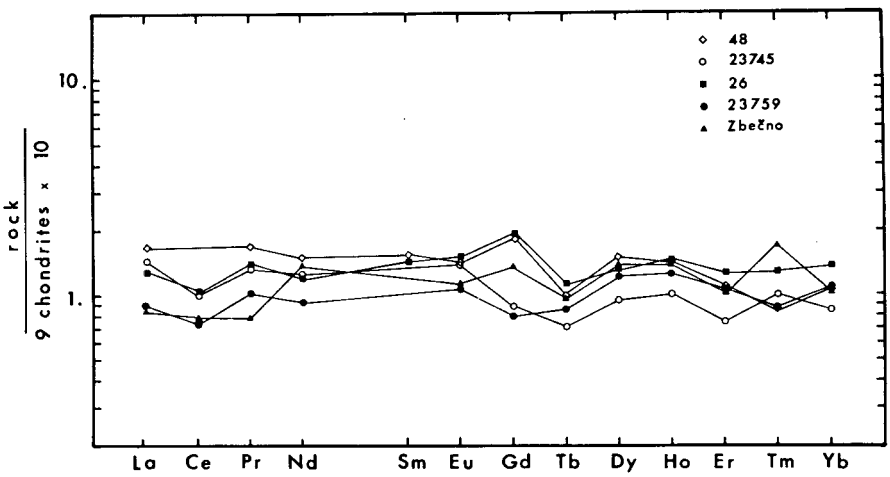

Fig. 5. Relative rare earth element patterns for rock from Macquarie Island [48,26], spilite from Upper Proterozoic of Czech Massif (Zbečno), and meta-tholeiites from Archaean greenstone belt – Norseman Western Australia.

Papua or those of Macquarie Ridge were derived from a mid-ocean ridge or pre-arc environment.

5. Significance of the island arc tholeiitic series

Rocks of the island arc tholeiitic series are at least as and perhaps more common than those of the calc-alkaline series in Tertiary oceanic island cores. These arcs are characterized either by Quaternary volcanism which is principally or entirely tholeiitic (i.e., Japan [13], New Britain, the southern Kurile and South Sandwich islands), by an extensive mid-Tertiary or older tholeiitic basement on which more recent calc-alkaline ± shoshonitic volcanism may have developed (i.e., Saipan-Guam, Fiji, the Solomon and Aleutian islands, and the Antilles), or both (Tonga). Only the abundance of Quaternary calc-alkaline volcanism in the Americas and Lesser Antilles lured petrologists into believing it was the dominant rock type of island arcs.

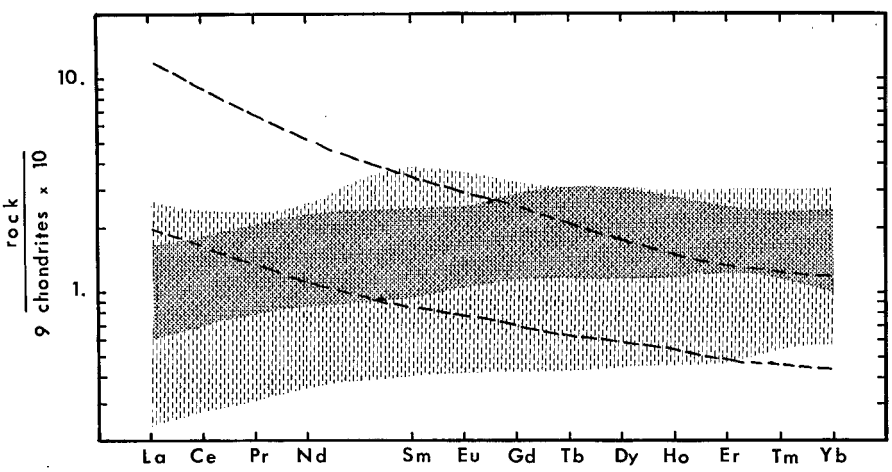

Fig. 6. Relative rare earth element patterns of published and new data on island arc tholeiites (vertical stiples), compared with patterns on mid-ocean tholeiites [22] and published and new data on calc-alkaline, shoshonitic and alkaline rocks of island arcs (between dashed lines).

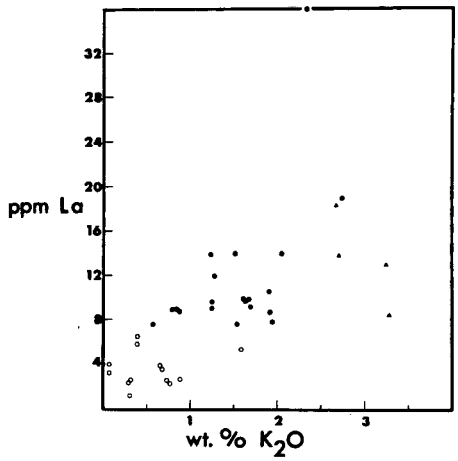

Fig. 7. Correlation of K_2O and La content for island arc rocks including published and new data. Open circles – island arcs tholeiites; full circles – calc-alkaline rocks; triangles, shoshonitic rocks from islands arcs. Data from [this paper, 8, 20, 25].

Although island arc basement volcanics are not well studied, they often bear evidence of submarine volcanism (pillow lavas associated with radiolarian cherts, pelagic limestones, and turbidites) and have an abundance of spilites, keratophyres, or entire ophiolite sequences including alpine ultramafics. As already stressed, an important corollary of recognizing the integral importance of non-calc-alkaline volcanism in island arcs is that the tholeiitic/ophiolitic rocks of

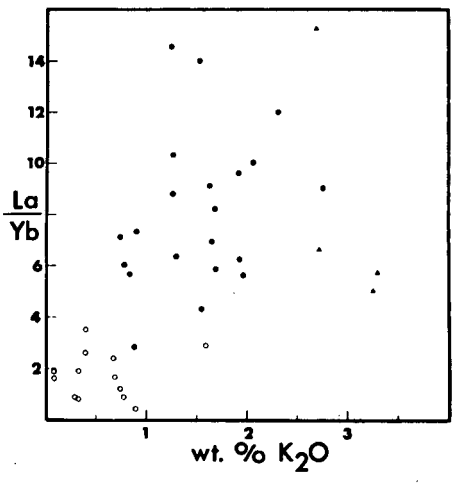

Fig. 8. Correlation of K_2O content and La/Yb ratio. Symbols and data as in fig. 7.

island arc basements or other orogenic belts *could* have originated where they now are, i.e., at sites where lithosphere is consumed, not created. While it may yet prove correct. it is at least unnecessary to assume [27] that alpine ultramafic or ophiolite complexes originated at mid-ocean ridges and were moved to their present location by sea-floor spreading.

The volumetric significance of this volcanic basement is difficult to assess but is likely to equal or exceed that of its calc-alkaline successors. For example, the so-called "non-volcanic ridges" of Indonesia [28] and Tonga [29], are known to include rocks of more tholeiitic than calc-alkaline affinities.

Where occurring alone, the spilites and keratophyres of this association are indistinguishable from rocks of the island arc tholeiitic series [8,9]. The mafic lavas of ophiolite complexes are usually tholeiitic [30], but whether any particular such rocks can be designated as mid-ocean ridge or island arc tholeiites depends on our ability to distinguish between these two.

6. Comparison between the mid-ocean and island arc tholeiitic series

Salient chemical features of island arc and abyssal tholeiites are summarized in table 1. In such distinctive aspects as Th, U, and REE contents and K/Rb, La/Yb, and Th/U ratios, they resemble each other more than any other terrestrial rock type. Rocks of both series have lower concentrations of K and related trace elements than do those of calc-alkaline or continental and Hawaiian tholeiitic series, but this is more pronounced in abyssal tholeiites.

It is important, however, to emphasize this and other differences. Although similar in Al_2O_3 and TiO_2 ranges, rocks of the island arc tholeiitic series have a greater silica range and many more rocks with $SiO_2 > 52\%$. At any silica content they have less FeO, MgO, Ni, and Cr but more K, Rb, Ba, Cs, Pb, and Sr and these differences appear critical. The contrast in iron enrichment pattern during fractionation is marked and illustrated in fig. 2.

Not only do island arc tholeiitic rocks have higher concentrations of alkali elements, these also have a somewhat more radiogenic character. When normalized to 0.7080 for the MIT Sr standard, initial

Sr 87/86 ratios of Pacific abyssal tholeiites average 0.7024 [31], whereas those of arc tholeiites average 0.7034 in Japan [32], 0.7038 in Saipan [33], 0.7035 in New Britain [34], 0.7041 in Fiji [8], and 0.7042 in the Antilles [9]. Hawaiian tholeiites have ratios of 0.703-0.704 [35]. Whether this variation reflects genetically significant differences or just lateral upper mantle inhomogeneities is not clear.

Both Japanese and Antillean tholeiites are slightly richer in Pb^{207} than their abyssal counterparts when corrected for instrumental mass fractionation [8, 32, 36–38]. In addition, Japanese tholeiites differ from abyssal tholeiites in having an inverse rather than positive correlation between Pb 206/204 and U238/Pb204 [37]. Thus, although neither Sr nor Pb isotopic compositions unambiguously delimit genetically significant differences between the two tholeiitic series, they may suggest a somewhat greater radiogenic component in the geochemical reservoir from which the arc variety came.

7. Genesis of the island arc tholeiitic series

Island arc tholeiites are chemically analogous to, but nevertheless different from, both calc-alkaline and abyssal tholeiitic rocks. Whatever is said about their genesis depends in part on which analogy is emphasized.

If they are but a variation on the traditional calc-alkaline theme, one must modify theories of calc-alkaline genesis to explain the chemical differences. They cannot be explained by fractionation under lower and less constant pO_2 from the same basaltic parent which under other conditions produced a calc-alkaline because: 1) this explains neither the difference in SiO_2 mode nor the differences in alkali element, REE, Th, or U concentrations which are persistent throughout the series and seem to include the basaltic members; 2) the abundance of intermediate and acidic rocks in both series makes fractionation of basaltic magma alone seem inadequate; and 3) it fails to explain the consistent temporal and spatial relationships between rocks of the two series. Calc-alkaline magmatism (s.c.) in oceanic island arcs may alternatively resutl from partial melting of amphibolite or quartz eclogite either at the base of an arc or in lithosphere thrust beneath it [39], or from the incongruent melting of enstatite in the upper mantle under high pH_2O [40]. Neither process is well enough understood to see how it could be modified to also consistently yield rocks of the island arc tholeiitic series at the right time and place.

If, instead, the similarities with abyssal tholeiites are emphasized, the residual differences might be explained by adding small amounts of radiogenic alkalis and oxygen to a tholeiitic magma or magma source. One possible source for both is the lithosphere thought to be thrust beneath island arcs [1], particularly if that lithosphere includes minor alkali basalts or has retained some portion of its sedimentary component and free water.

Oceanic sediments are rich in K, Rb, Ba, Cs, Pb, and Sr [41], and their Sr and Pb are enriched in radiogenic isotopes [42]. One would expect some fraction of these volatile elements together with H_2O to be lost either through dehydration of sinking lithosphere (the upper portion of which is likely to move from greenschist to amphibolite or eclogite facies assemblages) or upon its initial melting. The low-melting fraction of this hydrated and dominantly mafic upper surface of lithosphere would be andesitic to dacitic in composition [39]. The depth at which this would be expected is dependent on the angle and rate of underthrusting and the composition of the slab. Using isotherms calculated by Minear and Toksoz [43], any such melt would rise into even hotter overlying upper mantle, probably resulting in its partial fusion and/or hydration. Such a process might yield olivine tholeiitic magma either near the surface of underthrusting or at depths of 15-35 km in a rising diapir of previously depleted but partially molten and/or hydrous peridotite [44].

Magma evolution in more oxygen-rich environments will result in less iron enrichment during fractionation [45]. Rocks of the island arc tholeiitic series seem to have had access to more oxygen than those of the abyssal tholeiitic series (fig. 2). If Cr^{+2} as well as Fe^{+2} is oxidized under such positions and precipitated as chromite, this could account for both the order of magnitude lower Cr contents in island arc tholeiitic basalts and the concentrations of podiform chromite deposits in island arcs [46]. Absence of inflection or negative Eu anomalies in island arc tholeiitic REE patterns suggests that olivine more than plagioclase has dominated early fractionation

history and the low Ni contents agree. This is also consistent with the presence of increased pO_2 [47]. All island arc tholeiitic basalts reported in the literature have too little MgO, Ni, and Cr to have been in equilibrium with peridotite. Therefore, they either represent fusion products from a mafic rather than ultramafic source (as discussed above), or have lost considerable olivine by crystal settling. The range of rock types and characteristic plagioclase phenocrysts of both the island arc tholeiitic and calc-alkaline series suggest that both have undergone extensive and probably near-surface fractionation.

If island arc tholeiitic rocks do result from mixing the early melting fraction of descending lithosphere with that of the upper mantle overlying it [37], this could explain both the continuity between the island arc tholeiitic and calc-alkaline series and the intimate association of tholeiitic and alpine ultramafic rocks in island arcs and other orogenic zones.

8. Conclusions

Is is possible and important to distinguish between calc-alkaline and tholeiitic rock series in island arcs and to recognize the volumetric significance of the latter. We have attempted to facilitate this distinction by delineating the general geochemical characteristics of the island arc tholeiitic series and its REE features in particular. There are similarities between abyssal and arc tholeiites but also significant differences.

The current eruption of such tholeiites in island arcs demonstrates conclusively that rocks which occur in arc basements and have similar chemical characteristics *could* have originated in an arc environment and need not be rinds of oceanic crust. This is not to say that some ophiolite complexes or their purely eruptive equivalents in island arcs or orogenic zones could not be the original oceanic crust upon which the arc developed or pieces of allochthonous crust moved into place by sea floor spreading. It merely argues that this is not necessarily so, as is often assumed. Whether any particular tholeiite or ophiolite complex now in an island arc or Alpine belt *did in fact* originate above descending lithosphere is a matter for debate, the resolution for which depends on our ability to distinguish between arc and abyssal tholeiites.

Whether one explains the genesis of the island arc tholeiitic series by reference to a mafic or ultramafic source depends on whether its similarities to the calc-alkaline or tholeiitic series are emphasized. It may be possible to explain their distinguishing chemical features by envisaging an imprecise mixture of partial melts from descending lithosphere, enriched in alkalis and water, and overlying upper mantle. Such a process may yield both the continuum between the calc-alkaline and island arc tholeiitic series and the association of the latter with alpine ultramafic rocks in island arcs.

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References

[1] B. Isacks, J. Oliver and L. Sykes, Seismology and New Global Tectonics, J. Geophys. Res. 73 (1968) 5858.
[2] M.A. Peacock, Classification of igneous rocks, J. Geol. 39 (1931) 54.
[3] S.R. Nockolds and R. Allen, The geochemistry of some igneous rock series, Geochim. Cosmochim. Acta 4 (1953) 105.
[4] S.R. Taylor, Trace element chemistry of andesites and associated calc-alkaline rocks, in: Proc. Andesite Conf. (ed. A.R. McBirney) (1969) 43.
[5] T. Tomita, Geological and petrological study of Dogo, Uki, Geol. Soc. Tokyo J. 39 (1932) 609.
H. Kuno, Origin of Cenozoic petrographic provinces of Japan and surrounding areas, Bull. Volcanol. 20 (1959) 37.
[6] H. Kuno, Lateral variation of basalt magma across continental margins and island arcs, Bull. Volcanol. 29 (1966) 195.
[7] P. Jakeš and A.J.R. White, Structure of the Melanesian Arcs and correlation with distribution of magma types, Tectonophysics 8 (1969) 223.

[8] J.B. Gill, Geochemistry of Viti Levu, Fiji and its evolution as an island arc, Contr. Mineral. Petrol. 27 (1970) 179.
[9] T.W. Donnelly, J.J.W. Rogers, P. Pushkar and R.L. Armstrong, Chemical evolution of the igneous rocks of the Eastern West Indies: an investigation of thorium, uranium and potassium distributions, and lead and strontium isotopic ratios (1970) (in press).
[10] P.E. Baker, Comparative volcanology and petrology of the Atlantic island-acrs, Bull. Volcanol. 32 (1968) 189.
[11] H. Kuno, Petrology of the Hakone Volcano and adjacent areas, Japan, Bull. Geol. Soc. Am. 61 (1950) 957.
[12] C.E. Tilley, Some aspects of magmatic evolution, Q. J. Geol. Soc. Lond. 106 (1950) 37-61, pp. 42 and 51.
[13] A. Sugimura, Spatial relations of basaltic magmas in island arcs, in: Basalts (eds. H.H. Hess and A. Poldevaart) 2 (1968) 537-571, figures 9 and 10.
[14] J.G. Fitton and D.J. Hughes, Volcanism and plate tectonics in the British Ordovician, Earth Planet. Sci. Letters 8 (1970) 223.
[15] A.J.R. White, P. Jakes and D.M. Christie, Composition of greenstones and the hypothesis of sea floor spreading in the Archaean, Proc. Conf. on Archaean Rocks, Perth, W.A. (1970) (in press).
[16] S.R. Taylor, Geochemical analysis by spark source mass spectrography, Geochim. Cosmochim. Acta 29 (1965) 1243.
[17] J. Schilling, Red Sea floor origin: rare-earth evidence, Science 165 (1969) 1357.
[18] A. Masuda, Lanthanides in basalts of Japan with three distinct types, Geochem. J. 1 (1966) 11.
A. Masuda, Geochemistry of lanthanides in basalts of Central Japan, Earth Planet. Sci. Letters 4 (1968) 284.
[19] F.A. Frey, M.A. Haskin, J.A. Poetz and L.A. Haskin, Rare earth abundances in some basic rocks, J. Geophys. Res. 73 (1968) 6085.
[20] S.R. Taylor, A.C. Capp, A.L. Graham and D.H. Blake, Trace element abundances in andesites. II. Siapan, Bougainville and Fiji, Contr. Mineral. and Petrol. 23 (1969) 1.
[21] A. Ewart and A. Graham, Personal communication (1970).
[22] P.W. Gast, Trace element fractionation and the origin of tholeiitic and alkaline magma types, Geochim. Cosmochim. Acta 32 (1968) 1057.
[23] R. Kay, N.J. Hubbard and P.W. Gast, Chemical characteristics and origin of oceanic ridge volcanic rocks, J. Geophys. Res. 75 (1970) 1585.
[24] J. Schilling and J.W. Winchester, Rare earth contribution to the origin of Hawaiian lavas, Contr. Mineral. Petrol. 23 (1969) 27.
N.J. Hubbard, A chemical comparison of oceanic ridge, Hawaiian tholeiitic and Hawaiian alkalic basalts, Earth Planet. Sci. Letters 5 (1969) 346.
[25] S.R. Taylor and A.J.R. White, Trace element abundances in andesites, Bull. Volcan. 29 (1966) 177.
[26] W.R. Dickinson and T. Hatherton, Andesitic volcanism and seismicity around the Pacific, Science 157 (1967) 801.
[27] R.S. Dietz, Alpine serpentinites as oceanic rind fragments, Bull. Geol. Soc. Am. 74 (1963) 947.
H.H. Hess, The oceanic crust, the upper mantle and the Mayaguez serpentinized peridotite. In: A study of serpentinite (ed. C.A. Burke) Natl. Acad. Sci. - Natl. Res. Council Publ. 1188 (1964) 169.
T.P. Thayer, Peridotite-gabbro complexes as keys to petrology of mid-oceanic ridges, Bull. Geol. Soc. Am. 80 (1969) 1515.
[28] K.S. Rudolfo, Bathymetry and marine geology of the Andaman Basin, and tectonic implications for southeast Asia, Bull. Geol. Soc. Am. 80 (1969) 1203.
[29] J.E. Hoffmeister, Geology of Eua, Tonga, Bernice P. Bishop Mus. Bull. 96 (1932).
A. Stewart, Personal communication (1970).
[30] E.M. Moores and F.J. Vine, The Troodos Massif, Cyprus and other ophiolites as oceanic crust, evaluation and implications, Phil. Trans. Roy. Soc. Lond. (in press).
H.L. Davies, Papuan ultramafic belt, 23rd Int. Geol. Congr., Prague, Repot., sec. 1 (1968) 209.
I.G. Gass, Is the Troodos Massif of Cyprus a fragment of Mesozoic ocean floor, Nature 220 (1968) 39.
[31] Z.E. Peterman, I.S.E. Carmichael and A.L. Smith, Sr 87/Sr 86 ratios of Quaternary lavas of the Cascade Range, Northern California, Bull. Geol. Soc. Am. 81 (1970) 311.
[32] C.E. Hedge and R.J. Knight, Lead and strontium isotopes in volcanic rocks from northern Honshu, Japan, Geochem. J. 3 (1969) 15.
[33] C.E. Hedge, Variations in radiogenic strontium found in volcanic rocks, J. Geophys. Res. 71 (1966) 6119.
Paul Pushkar, Strontium isotope ratios in volcanic rocks of three island arc areas, J. Geophys. Res. 73 (1968) 2701.
[34] Z.E. Peterman, G.G. Lowder and I.S.E. Carmichael, Sr 87/Sr 86 ratios of the Talasea series, New Britain, Territory of New Guinea, Bull. Geol. Soc. Am. 81 (1970) 39.
[35] J.L. Powell and S.E. DeLong, Isotopic composition of strontium in volcanic rocks from Oahu, Science 153 (1966) 1239.
[36] M. Tatsumoto, Isotopic composition of lead in volcanic rocks from Hawaii, Iwo Jima and Japan, J. Geophys. Res. 71 (1966) 1721.
[37] M. Tatsumoto, Lead isotopes in volcanic rocks and possible ocean floor thrusting beneath island arcs, Earth Planet. Sci. Letters 6 (1969) 369.
[38] J.A. Cooper and J.R. Richards, Lead isotopes and volcanic magmas, Earth. Planet. Sci. Letters 1 (1966) 259.
[39] T.H. Green and A.E. Ringwood, Genesis of the calc-alkaline igneous rock suite, Contr. Mineral. Petrol. 10 (1968) 105.
[40] H.S. Yoder Jr., Calc-alkalic andesites: experimental data bearing on the origin of their assumed characteristics. In: Proc. Andesite Conf. (ed. A.R. McBirney) (1969) 77.
[41] K.K. Turekian and K.H. Wedepohl, Distribution of the elements in some major units of the earth's crust, Bull. Geol. Soc. Am. 72 (1961) 175.

G. Thompson, Analyses of B, Ga, Rb, and K in two deepsea sediment cores, Marine Geol. 6 (1968) 463.

[42] E.J. Dasch, F.A. Hills and K.K. Turekian, Strontium isotopes in deep-sea sediments, Science 153 (1966) 295. T.J. Chow and C.C. Patterson, The occurrence and significance of lead isotopes in pelagic sediments, Geochim. Cosmochim. Acta 26 (1962) 263.

[43] J.W. Minear and M. Nafi Töksöz, Thermal regime of a downgoing slab and New Global Tectonics, J. Geophys. Res. 75 (1970) 1397.

[44] D.H. Green and A.E. Ringwood, The genesis of basaltic magmas, Contr. Mineral. Petrol. 15 (1967) 103. D.H. Green, Composition of basaltic magmas as indicators of conditions of origin: application to oceanic volcanism, Phil. Trans. Roy. Soc. London (1970).

[45] P.L. Roedder and E.F. Osborn, Experimental data for the system MgO-FeO-Fe_2O_3-$CaAl_2Si_2O_8$-SiO_2 and their petrologic implications, Am. J. Sci. 264 (1966) 428. D.L. Hamilton and G.M. Anderson, Effects of water and oxygen pressures on the crystallization of basaltic magmas. In: Basalts (ed. H.H. Hess and A. Poldevaart) vol. 1 (1968) 445.

[46] T.P. Thayer, Gravity differentiation and magmatic reemplacement of podiform chromite deposits. In: Symposium on magmatic deposits (ed. H.D.B. Wilson) (1969) 132. Econ. Geol. Mon. 4.

[47] I. Kushiro, H.S. Yoder and M. Nishikawa, Effect of water on the melting of enstatite, Geol. Soc. Am. Bull. 79 (1968) 1685.

[48] F. Chayes, The chemical composition of Cenozoic andesite, in: Proc. Andesite Conf. (ed. A.R. McBirney) (1969) 1.

[49] K.S. Heier and J.J.W. Rogers, Radiometric determination of thorium. uranium and potassium in basalts and in two magmatic differentiation series, Geochim. Cosmochim. Acta 27 (1963) 137.

[50] A.E.J. Engel, C.G. Engel and R.G. Havens, Chemical characteristics of oceanic basalts and the upper mantle, Bull. Geol. Soc. Am. 76 (1965) 719.

[51] J.A. Philpotts, C.C. Schnetzler and S.R. Hart, Submarine basalts; some K, Rb, Sr, Ba, rare-earth, H_2O, and CO_2 data bearing on their alteration, modification by plagioclase, and possible source materials, Earth Planet. Sci. Letters 7 (1969) 293.

[52] S.R. Hart, K, Rb, Cs contents and K/Rb, K/Cs ratios of fresh and altered submarine basalts, Earth Planet. Sci. Letters 6 (1969) 295.

[53] F. Aumento, Diorites from the Mid-Atlantic Ridge at 45°N, Science 165 (1969) 1112.

[54] I.S.E. Carmichael, The petrology of Thingmuli, a Tertiary volcano in eastern Iceland, J. Petrology 5 (1964) 435.

[55] L.R. Wager, The major element variation of the layered series of the Skaergaard intrusion and a re-estimation of the average composition of the hidden layered series and of the successive residual magmas, J. Petrology 1 (1960) 364.

The authors wish to make the following corrections :

p. 17 Abstract: line 4. "... rocks with these characteristics occur above shallower earthquake foci ...".

p. 17 Add to bottom of column two "and alkali basalts or shoshonites afterwards".

p. 20 Column two, 1st line of 3rd paragraph under Results. "Fig. 6 summarizes published island arc REE data ...".

p. 24 Column two, line 4. "... it may yet prove correct, it is at least unnecessary to ..."

p. 25 Column one, 7th line of 2nd paragraph under Genesis. "... calc-alkaline series because ..."

p. 25 Lines 16-17 of above paragraph. "Calc-alkaline magmatism (s.s.) in oceanic island arcs may alternatively result from partial melting of ..."